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CASCADE ANDESITES OF OREGON

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ABSTRACT

This paper is a summary of studies made on extensive samples of the main bulk of the Cascan formation of the Cascade Mountains of Oregon. The rocks show many mineral combinations, yet all minerals show that they were once a part of a common magma, altered by incorporated rocks, that was moving toward a fixed mineral objective and stopped in this march at various places. Hence, the various mineral assemblages are only one frame in a cinema-like picture, the final scene of which would have shown all the rocks to be composed of pigeonite and labradorite. The effects of magma contamination, resorption, reaction, and arrested stages are found to be misleading when isolated flows or areas are examined. However, detailed optical studies on many hundreds of selections from many localities resolve the varied and complex results into a simple, straightforward change from a dacitic magma toward a basic andesite end product.

INTRODUCTION

The Cascade andesites are part of the Cascan formation which has been defined by Hodge as an accumulation of lavas with some interbedded pyroclastics that lies with no unconformity on the Madras, The Dalles, Rhododendron, and Troutdale formations. In fact they are the upper part of the above formations, where composed dominately of lavas. The Cascan formation lies below the lavas of the younger volcanoes, such as Mount Hood and Mount Jefferson, and the many other still younger lavas. It is thought to be of early late Pliocene or early Pleistocene age.¹

The rocks described are samples that were collected between the years 1924 and 1932 in that portion of the central Cascade Mountains of Oregon, extending from the Columbia River to Diamond Peak. Several thousand specimens were collected and several hundred thin sections prepared therefrom and studied.

CLASSIFICATION

Many attempts were made to classify hand specimens of these rocks according to their texture, color, and mineral content. All such attempts

¹ Hodge E. T., Geology of the lower Columbia River: *Bull. Geol. Soc. Am.*, **49**, 831-930 (1938).

proved futile when it was found that the composition of rocks of similar texture were not alike mineralogically.

Thin section studies using ordinary methods appeared for a time to indicate a possible classification based upon the presence of olivine, hypersthene, augite, hornblende, and certain plagioclases. However, continued study proved that many of the above minerals are variable in their character and their reaction and absorption effects indicate that many of the minerals are present only as part time participants in the history of a rock that has undergone a progressive change. These changes carried some rocks but a little way, others farther, and still others completely toward their destination. Even the plagioclases are found to be so variable, both within each rock and in suites of rocks, that only a few specimens can be placed in a group that have nearly identical plagioclases. The augites, at first thought to characterize the rocks, were found to be rare and most so-called "augites" are pigeonites.

It is the opinion of the writers that Cascan rocks, though of various external appearances and containing different minerals and giving different chemical analyses, are, as a whole, definable in and members of one group.

TEXTURE

Ninety per cent of the Cascade andesites are of porphyritic texture and the grains show a large range in size. The larger crystals of one specimen may be a third or a fifth as large as the largest crystals in another specimen. The phenocrysts are not wholly feldspars, but in some cases consist of olivine, fayalite, hornblende, and pyroxene. A few of the rocks contain a large percentage, but most of them have only a small percentage of glass. The rocks with much glass have hyalophitic, hyalopilitic, and hyalotrachytic textures. The other rocks are diabasic, ophitic, and intergranular in texture.

MINERALS

Feldspars

The feldspars of the Cascade andesites range in amount from 55 to 65 per cent in the rocks where visible glass in the hand specimen is not apparent. Even the glassy rocks show a few crystals of feldspar. The amount of glass varies inversely with the amount of feldspar.

The feldspars are all plagioclases including, perhaps, anemousite. No albite, anorthite, nor orthoclase is found. The chemical analyses show a range from 0.28 to 0.48 of a per cent of potassia which is probably present in the intermediate plagioclases or in the hornblendes. Each individual rock exhibits a wide range in the kind of plagioclase present. It is common to find in a rock bytownite phenocrysts, laths and smaller phenocrysts of labradorite, and crystallites of andesine.

The plagioclase grains are so variable in size that it is improper to class them as of the "first and second generation," or by any division of sizes. The sizes vary both in each specimen and in different specimens of rocks of the same composition. A difference in the size of only one hundredth of a millimeter is commonly found in a gradational series from the largest to the smallest crystal. Thus, phenocrysts several millimeters in length may be the largest crystals in a rock where the size grades by small amounts down to the microlites. Phenocrysts as large as 10 millimeters are frequently seen, but the average is from 2.5 millimeters to 4.5 millimeters. The rocks as a whole are fine grained and range in size in accordance with their time of crystallization.

The large, early formed crystals are euhedral or subhedral. The crystals of average size are usually lath-shaped, elongated parallel with the a -axis, and have sharp, often jagged, and irregular ends. The crystallites and microlites are also generally elongated parallel to the a -axis.

The plagioclases contain numerous inclusions of "magnetite," bits of glass, bubbles, and small globules of solidified ground mass. Inclusions rarely found are olivine, fayalite, augite, pigeonite, and a few apatites. Plagioclase microlites are included in hornblende, augite, and in the large plagioclases. The microlites included in the plagioclases are of identical composition to the plagioclase microlites in the groundmass.

Plagioclase laths are oriented around the hypersthene crystals and tiny plagioclases are often intermingled with masses of pigeonite grains. No fractured plagioclases are found except in a few flows located on the slopes of Mount Hood, and in some late flows on other volcanoes. These flows are made of a meshwork of plagioclases, sealed together by glass, from which the excess liquid drained before freezing.

The albite type of twinning is found in nearly all crystals. Carlsbad twinning is also common and in some large crystals the Carlsbad twins occur unassociated with albite twins. Combined albite and Carlsbad twins are so common that sections of them were used for determinative purposes. Baveno, Manebach, and pericline twins are rarely found. The " A " type of Carlsbad twin is more common than the " B " as determined by a Fedorov universal stage. The medium-sized crystals usually have albite or combined albite and Carlsbad twinning, occasionally Carlsbad " A ," but no " B " twins, and rarely Baveno, Manebach, and pericline twins. Some of the larger crystallites have albite twins and a few have Carlsbad twins. One member of the albite twins is commonly much larger. The smaller crystallites and microlites usually show a simple "needle." The contacts of the zones are seldom sharp and some crystals show a gradual change in composition from the center to the periphery. No oscillatory zoning is found.

Determination of the optic sign and angle by means of a universal stage on sections of zoned crystals cut absolutely, or nearly perpendicular to the optic axis, showed that the inner zones in the larger crystals are calcic and the outer zones are successively more sodic in composition. As an example, the central core in a crystal was found to be bytownite of composition $\text{Ab}_{26}\text{An}_{74}$, with an optic angle of 87° and the optic sign was negative. The next peripheral zone was found to be labradorite of composition $\text{Ab}_{32}\text{An}_{68}$ and the optic sign had changed to positive with $2V=82^\circ$. The third outer zone was also labradorite, but with the more sodic composition $\text{Ab}_{40}\text{An}_{60}$. The sign was still positive, but $2V$ had decreased to 77° . The outermost zone had the composition $\text{Ab}_{50-52}\text{An}_{50-48}$. It was about midway between andesine and labradorite. The sign was positive with $2V=78^\circ$. Although the optic angle varied only 1° from the adjoining zone, the extinction showed it to be more sodic.

The kind of zones shown by our crystals depend upon their size and range in composition from bytownite to andesine, or from labradorite and to andesine or basic oligoclase. The crystallites are always more sodic than the larger grains; thus, if the phenocrysts are andesine, the crystallites may be oligoclase of the composition $\text{Ab}_{35}\text{An}_{15}$.

Absorption has affected the larger and more calcic crystals. Some embayments so formed are filled with groundmass or with more sodic plagioclase. No sodic plagioclases have embayments, but a few have a slightly dissolved periphery. Passages are dissolved through the sodic rims into the cores which are more or less replaced by sodic plagioclase.

Unknown Mineral

There are areas in the groundmass of the andesites that do not have crystal form but show birefringence. These masses are sometimes colorless, but usually are of a milky white color. They have no cleavages and fractures pass from the surrounding glass through them and into the glass again. There is often no visible contact between these areas and the adjacent glass, and they appear to be transitional into the glass. Grains of magnetite are imbedded in them.

Their general appearance is that of quartz or tridymite and they have approximately the right index of refraction for this supposition. The masses are so small that a Johannsen auxiliary lens was used to obtain a biaxial figure with an angle of about 90° . Some of the areas are optically positive and others are negative. The optic angle is too large and the birefringence too high for tridymite.

The presence of silica in these rocks if not always in excess, is present in amounts sufficient to react with the orthosilicates, and this is considered evidence that the unknown mineral is not nepheline. Another pos-

sibility is that it may be anemousite² which has indices of 1.53–1.55, a positive optical sign and $2V = 10^\circ$ – 85° . The indices of the unknown mineral are in close agreement and $2V$ is between 78° and 90° for these masses which are optically positive, and 84° to 90° for those which are optically negative.

Another possibility is that material was assembled from the residual magma, ready to crystallize into a feldspar, but was caught in this initial stage and frozen in with the glass. If such is the case, then the material may be oligoclase. At the composition $\text{Ab}_{84}\text{An}_{16}$ the optic sign of oligoclase changes from positive to negative. The index of refraction and the birefringence are those of oligoclase.

Olivine

The name olivine here is used for chrysolite, hyalosiderite, hortonolite, and fayalite. The olivine grains form a gradational series of sizes with, however, a large number of grains occurring in a limited range of sizes. Casual inspection gives the impression of two distinct generations of olivine grains, which is true of only few rocks. The evidence suggests that olivine was once present in all rocks, but has been lost through resorption. Small olivine grains occur as inclusions in plagioclase.

The larger olivines are generally euhedral, a form which rapidly is lost as the size decreases so that some of the smaller crystals have rarely more than a single pinacoid or dome. Most of the smaller crystals are globular forms conforming to the adjacent grains against which they grew.

No twinned crystals nor zonal growths of olivine are observed. Only one direction of cleavage is noted and this, except in rare cases, is poorly developed. The optical properties are gradational from fosterite to fayalite. One exception is the constant concurrence of the vibration B(Y) direction parallel to the crystallographic *c*-axis and axis of elongation, in all crystals where determined.

The mean index ranges from 1.65 to $1.82 \pm .03$ and the average of all determinations is 1.75. However, the determinations of the indices above 1.75 are not as consistent as those below and are not given equal approval.

Birefringence determinations with a Berek compensator are more consistent than those for index and greater dependence is placed upon them. In unaltered crystals the birefringence ranges from 0.035 to 0.049. Weathered or altered olivines have a higher birefringence, as high as 0.056 in one specimen.

In many crystals the interference is of little help in locating the acute

² Barth, T. F. W., Pacificite, an anemousite basalt: *Jour. Wash. Acad. Sci.*, **20**, No. 4, 60–68 (1930). Petrography of the Pacific lavas: *Am. Jour. Sci.*, **21**, 401–402 (1931).

bisectrix. The sections used for this determination were carefully selected and those determinations that gave over $\pm 3^\circ$ variations were rejected. The optic angle ranges in size from $(+)86^\circ$ to $(-)50^\circ$. Only one crystal was measured with a $(-)50^\circ$ optic angle; most of the crystals have an optic angle of more than $(-)60^\circ$. Most of the olivine grains show a negative sign; a few, a positive sign. In those crystals with an optic angle near 90° the dispersion is weak, apparently $v > r$; the dispersion $r > v$ becomes more distinct with the decrease of the optic angle below 80° .

The few positive results from many crystals tested are plotted in Fig. 1, in which the upper curve shows the optic angle and the lower curve the maximum birefringence, after the manner of Winchell.³ Comparing results it appears that crystals with a birefringence below 0.044 are in close agreement, but the birefringent curve above 0.044 rises above Winchell's plot. Hence, the olivine in these rocks ranges from fosterite through chrysolite ($2V = 86^\circ - 88^\circ$), hyalosiderite, and hortonolite, almost to fayalite. The manganic oxide in the analysis is sufficient for hortonolite.

The optical data implies the chemical range of the olivine to be from 98 Fo-2 Fa to 6 Fo-94 Fa. Following Winchell's figures and textual data on pages 166-168, our data indicate the presence of fosterite in these rocks which is doubtful in view of the usual occurrence of fosterite. Also the crystals studied show only one set of poorly developed cleavages, whereas true fosterite is generally described as having good (010) and (001) cleavages. The least index of refraction for fosterite is (Np) 1.635, whereas in our crystals the index is below 1.64 for this direction of vibration. Further, the birefringence for every crystal examined, except one with 0.035, is too high; ranging from 0.037 to above 0.0384. One crystal with $(+)2V = 86^\circ$ has a birefringence of 0.037. The chemical analysis shows the grains contain a large amount of the fosterite molecule.

One interesting fact in each rock is the general concordance of grain size of the olivine with the chemical and optical character. In all cases, the phenocrysts and larger crystals are so rich in fosterite that they never fall below chrysolite; the medium sized grains are most abundant and range between chrysolite and hyalosiderite. The small grains are hortonolite. Very small groundmass granules of fayalite containing optically only 6 per cent Fo were found in one rock.

Between hortonolite (Fo23) and fayalite (Fa6) we found no other minerals. It is our opinion that the above break is not due to incomplete sampling.

Fosteritic olivine is unstable in the presence of silica, while fayalitic

³ Winchell, A. N., *Elements of Optical Mineralogy*, Part II, John Wiley & Sons, 168 (1927).

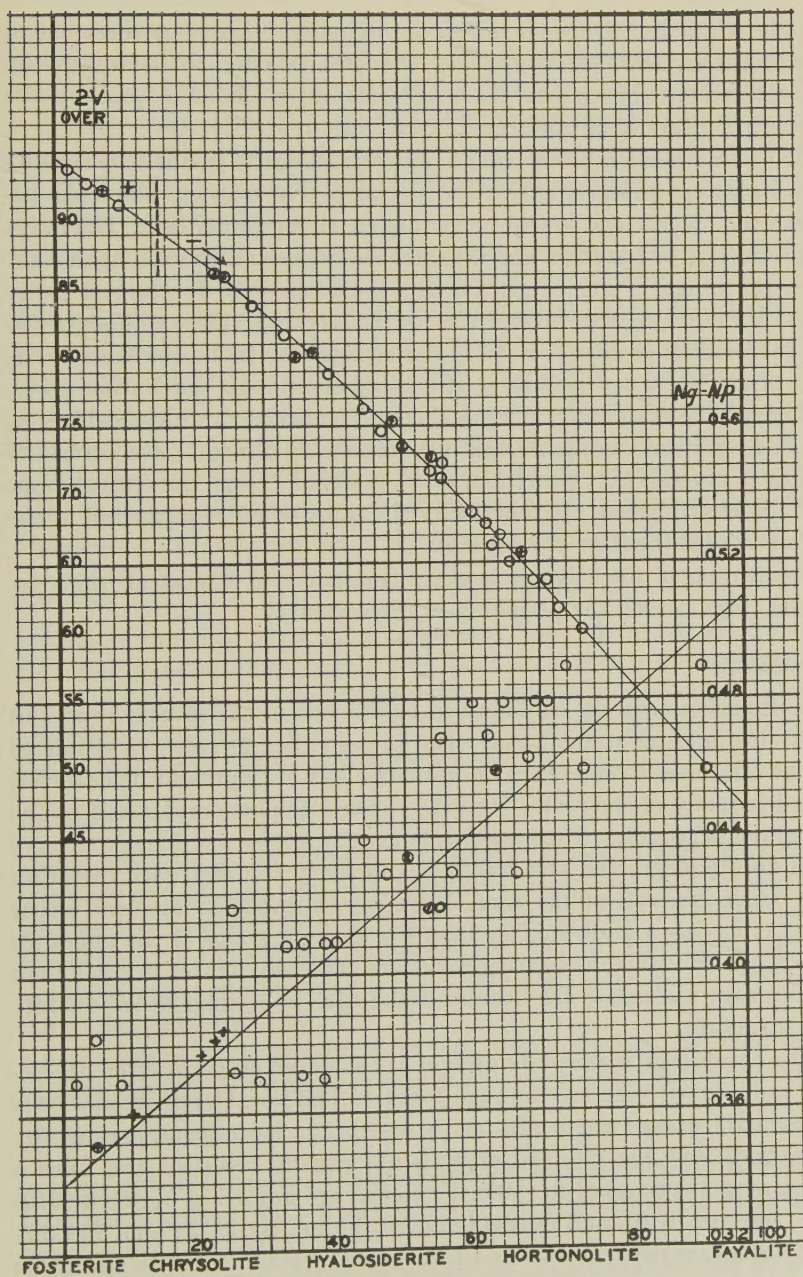


FIG. 1. The properties of the olivines.

olivine is stable. Hence, fayalite, if originally present in these lavas, would persist through absorption changes. A part of the magnetite crystallized late; hence iron was in solution and available to form fayalite grains in the late stage of olivine crystallization.

The same portion of the range between fosteritic grains and fayalite is not always shown by each rock specimen. The average difference between the highest and lowest percentage is about 15 per cent. The range in percentage of fosteritic and fayalitic molecules is always in agreement with the range in size of grains. The specimen with the largest range (Smock Prairie) showed Fo34–Fa56 for the largest grains, and Fo34–Fa66 for the smallest grains. T. F. W. Barth⁴ (page 380) found the same variation of gradation between large and small crystals in the Pacific lavas.

Though the grains vary in optical character and composition with size, they do not show zoning, though it is admittedly difficult to be certain of zoning in crystals that have been much resorbed. Thus, the optical angle and the birefringence are practically the same whether measured in the center or near the edge of a crystal.

Resorption of the olivines is quite common and almost all show reaction rings around the periphery of the crystals.

Hornblende

Hornblende crystals are found in only a few of the several thousand Cascade andesites studied, and when present they are always visible to the unaided eye. Evidence that hornblende was once present, is however, found in many of the rocks. It always occurs as phenocrysts and not as groundmass granules. The hornblende usually appears superficially to have been deeply weathered and stands out in sharp contrast with the other very fresh minerals of the rock. Moreover, hornblende does not occur in rocks having a grain smaller than the average grains of the andesites as a whole. Hornblende-bearing rocks are generally lighter in color than the average andesites, are usually trachytic, and some show definite flow structure.

The hornblende is generally present as elongated crystals, often needle-like and aligned with any flowage texture. The hornblende phenocrysts are euhedral except where affected by resorption. Hornblende is generally untwinned, but the twinned crystals are usually resorbed and have a twinning plane parallel to (100). The cleavage angle lies between (110) and (100), or (010), and ranges between the limits $124^{\circ}24'$ and $124^{\circ}30'$. The indices are between 1.64 and 1.66 for all crystals found except in one crystal which was between 1.66 and 1.68.

⁴ *Op. Cit.*

The measured birefringence of unaltered hornblende is 0.030 to 0.036 and the smallest crystals give the highest birefringence. The minima obtained are consistently about 0.010 with a variation of not more than 2 in their third decimal place. The extinction angle between Z and c ranged from 3° to 20°, but most of the crystals varied between 5° and 12°. The larger crystals have the larger extinction angles and the smaller "needles" have the smaller.

The pleochroism is strong and consistent in kind in all hornblendes as follows: X is yellow to yellowish-brown; Y is yellowish-brown to pale brown, and is always darker than the X colors; and Z is a dark brown to a brownish-black. The optic plane is parallel to (010) where determinable. The elongation is always positive. Dispersion is $r < v$ and inclined at a low angle. The optic angle varies between 74° and 83°, and the optic sign is negative. The crystals are so altered that determination of the optic planes and dispersion must be confined to a few crystals. The variations in optical properties are probably due more to temperature effects than to original variation in the chemical composition of the hornblende.⁵

Pyroxene

Next to the plagioclase the pyroxenes, chiefly pigeonite, are the most abundant minerals in these rocks. Pyroxene is absent in a few very glassy (80% or more glass) rocks. Monoclinic pyroxene is more abundant than orthorhombic. Enstatite was found in only two specimens; in all others the orthorhombic pyroxene is hypersthene.

Enstatite

Enstatite occurs as euhedral, prismatic crystals ranging from 2.5 mm. long and 1.0 mm. wide to 1.0 mm. by .5 mm., except in one specimen with an optic angle of $2V = 85^\circ$ to 87° . The optic angle is $2V = 72^\circ$ to 74° . Pleochroism is absent; birefringence is less than .01 and is generally less than .008; sign and elongation is positive and the mean index of refraction ranges between 1.67 and 1.68.

Dust grains of magnetite form the only inclusions in some of the smaller crystals. Brownish-red iron stains the edges of some crystals.

Hypersthene

Hypersthene occurs only as prismatic phenocrysts. Though of common occurrence, it never makes up more than ten per cent of the rocks and is never as abundant as pigeonite.

The crystals are euhedral or nearly so. The cleavages are generally

⁵ Winchell, A. N., Studies in the amphibole group; *Am. Jour. Sci.*, **107**, 309 (1924).

well developed and vary from $90^{\circ}30'$ to 92° . Distinct parting is common. The hypersthene is elongated parallel to the c -axis. The sizes range from 2.5 mm. to .75 mm. in length and from .85 mm. to .3 mm. in width.

The prismatic crystals are aligned with their long direction parallel to the flow structure of the rock. The plagioclase crystals are oriented in like manner around the hypersthene. The hypersthene contains many inclusions. Magnetite is the most abundant included mineral; globules of groundmass containing magnetite dust and plagioclase crystallites; shards of glass; and drops of liquid are also abundantly included.

The hypersthene is distinctly pleochroic; X gives a clear red or brownish-red, Y gives a yellowish-brown to yellowish-red; Z gives a pale light green in all cases except a few which show only a faint tinge of green. The index of refraction is variable from crystal to crystal even within the same rock. The limits for γ are 1.68 to 1.74.

The birefringence (Ng-Np) varies between 0.010 and 0.016. The optic sign is always negative and the optic angle varies between 57° and 87° . The optic plane is parallel to (010) in all crystals in which it is possible to determine the orientation. Elongation is positive and the dispersion is $r > v$, but never very strong.

The wide variation in optical properties in crystals within the same specimen varies with the sizes. The largest phenocrysts have consistently the largest optic angle, mostly between 83° and 87° with a birefringence from 0.010 to 0.012, and γ about 1.68 to 1.69. The smaller phenocrysts have a smaller optic angle, but higher index and birefringence.

The outer portions of the crystals are light gray when the center of a section is at extinction. The center of one crystal gave $2V = 85^{\circ}$, the edge $2V = 61\frac{1}{2}^{\circ}$, and between the value is $2V = 71^{\circ}$.

The hypersthene has been much absorbed by reaction and the resorbed areas are filled with augite, plagioclase, pigeonite, magnetite, and glass. The hypersthene weathers to a limonitic mass, brown or reddish-brown in color. Alteration worked in along the cleavages and fractures and eventually altered the entire crystal. The fine groundmass surrounding the crystal also becomes stained with the brown iron coloring.

In a very few specimens, the weathering has altered the hypersthene to a deep olive-green or, rarely, a yellowish-green bastite. Serpentine and fibrous bastite may form in the same specimen, but have never been found together in the same crystal.

Monoclinic Pyroxenes

The monoclinic pyroxenes are more abundant than the orthorhombic ones. Pigeonite is much more abundant than augite, and aegirine-augite has been found in only one specimen.

Aegirine-augite

Aegirine-augite occurs in a dike in the Rhododendron formation with a few crystals of hypersthene, the only other pyroxene in the rock. The crystals are all euhedral with a maximum size of 1.0 mm. by .45 mm. They are badly fractured and cleaved and a little absorbed. Lamellar twinning is quite common with (100) as the composition plane. Except for a few small grains of magnetite, the crystals are free from inclusions.

The aegirine-augite is a green or greenish-brown with distinct but not very strong pleochroism—X=pale green, Y=yellowish-green, and Z=light greenish-yellow to greenish-brown.

The birefringence is 0.029 and $Z \wedge c$ is $+72^\circ$. The optic sign is positive; $2V=94^\circ \pm 1^\circ$ and is constant for all the crystals measured in this specimen. Dispersion is strong and inclined with $r < v$.

Augite

Augite has been noted for a long time, in fact, described as the only monoclinic pyroxene in Cascade lavas, though our studies show it to be very subordinate to pigeonite in abundance. No reference to pigeonite in the Cascade lavas has been found. Barth⁶ (pages 383–391), found pigeonite to be common in the lavas of the Pacific, and Tsuboi⁷ found the same to be true in the Japanese volcanic rocks. The pyroxenes in our rocks are more like those in Japanese rocks than those of the Pacific region.

Augite forms almost always the larger of the monoclinic pyroxene crystals, and is often found as phenocrysts. The smaller crystals are nearly all pigeonite. It is often found inclosed by or moulded against plagioclase. Augite does not occur as a reaction product of olivine.

Phenocrysts of augite are usually euhedral or subeuhedral. The maximum size is 1.5 mm. long by 1.0 mm. in width. The crystals are stubby and seldom elongated parallel to the c -axis. The smaller augites, especially those inter-spaced between the plagioclase, often do not show a single crystal face.

One cleavage, at least, is well developed and intersects in (001) sections at angles from 87° to $88^\circ 30'$. The cleavage angles near the center of a crystal are usually constant, but may vary about a degree in a single section.

Twinning is rare and the composition plane may be (100) or (110), or rarely (001). If (110), the crystals will be geniculated with both twin

⁶ *Op. Cit.*

⁷ Tsuboi, Seitaro, Course of crystallization of pyroxenes from rock magmas: *Jap. Jour Geol. and Geo.*, **X**, Nos. 1–2, 61–82 (1932).

members, about equally developed. Lamellar twinning is with the base (001).

Inclusions of magnetite granules, feldspar crystals, shards and globules of glass, bubbles of liquid, and empty cavities are found. The inclusions have no definite arrangement though "rings" or zones of magnetite granules are found in euhedral crystals. The optical properties vary from crystal to crystal even within the same specimen.

The thin section color is usually a pale, light green to colorless, or less commonly a greenish-yellow and rarely a greenish-brown. The darker varieties and especially the smaller crystals have faint pleochroism: X is usually greenish, rarely greenish-yellow; Y is greenish-yellow, but usually more intense than X; Z is usually greenish-yellow of about the same intensity as X; and Z will rarely show a yellowish-brown or yellowish-red.

The index γ usually lies between 1.71 and 1.72. The index of the larger phenocrysts is invariable, while the indices of the smaller crystals vary between 1.72 and 1.735, and several tiny granules in the groundmass give 1.74.

The birefringence varies between 0.025 and 0.030 with an average of 0.027. The axial dispersion, $r > v$, is in most cases quite distinct, varies but a little in strength, and is always inclined. The extinction angle $Z \wedge c$ varies between $+35^\circ$ and $+52^\circ$, mostly between 41° and 45° .

The axial angle is perhaps more constant than any other optical property. The sign is always positive and $2V$ for all specimens, except one, falls between the limits 52° and 58° . One crystal has $2V = 62^\circ$. This is the one optical property that definitely separates augite from pigeonite. See Fig. 2. The base for this figure was taken from Winchell,⁸ p. 182, and the extinction angle $Z \wedge c$, and the birefringence plotted. The birefringence is quite constant throughout the series and the average is almost a horizontal line. The extinction angle $Z \wedge c$ slopes upward and to the right as does the curve for the optic angle. The two curves, $2V$ and $Z \wedge c$, appear in agreement. Those crystals having the low values for $2V$ have the smallest extinction angles. The plot shows a distinct break between the augite and pigeonite series and no crystals falling in the intermediate zone are found.

Augite, where affected, is only partially absorbed; deep embayments are of rare occurrence. The reaction products are pigeonite, magnetite, and tridymite is sometimes present in the reaction rings as small granules.

⁸ *Op. Cit.*

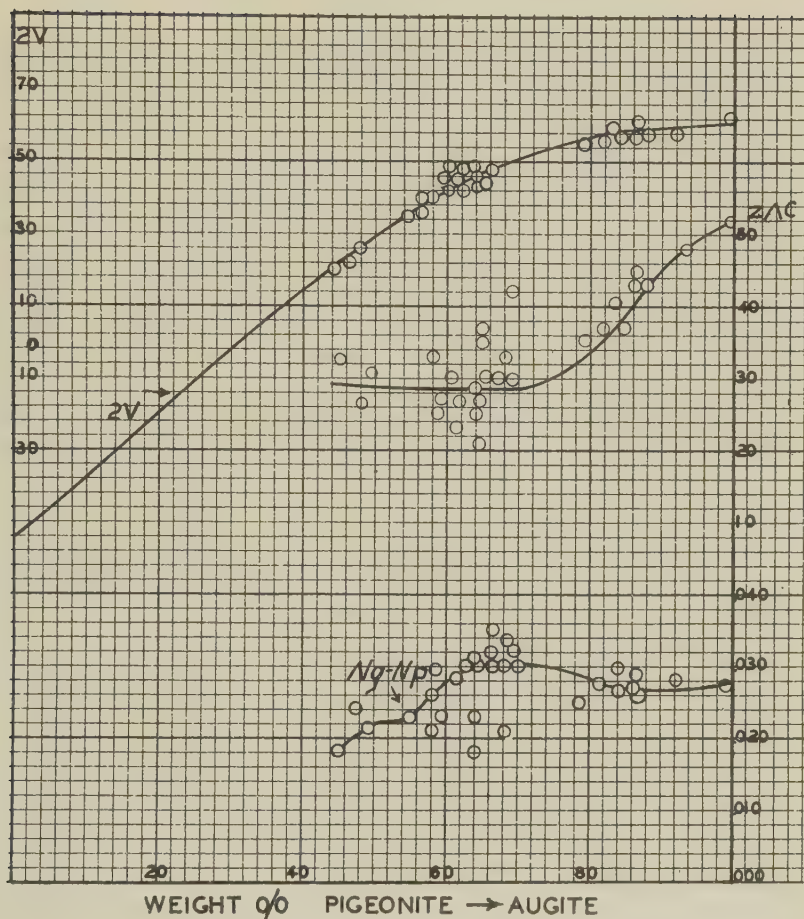


FIG. 2. The Pigeonite-Augite Relationship

Pigeonite

The amount of pigeonite increases as the amount of olivine is reduced. Pigeonite is very much more abundant than augite. In fact, only the larger pyroxenes are augite and all the smaller pyroxene crystals and granules in the groundmass are pigeonite. It occurs rarely as inclusions in the plagioclases and only in the outer borders of the larger crystals.

Pigeonite occurs usually in anhedral crystals and irregular masses that lie between the laths of plagioclase in the groundmass, especially in rocks with an ophitic texture. Large euhedral crystals of pigeonite form thick prisms, slightly elongated on the *c*-axis. Twinning occurs more frequently with pigeonite than augite. Even in the irregular masses, the

composition plane is sharp and good twinning is shown. Lamellar twinning on the face (100) was definitely determined in a few crystals; some appeared to be twinned on (110). Cleavage is irregular, and the average intersection of the cleavages is 91° with variations of several degrees. Parting is sometimes well developed parallel to (100).

Pigeonite in thin section is usually some shade of light green, sometimes greenish-brown, and rarely a reddish-brown. Pleochroism is faint, variable but distinct; Z and X almost invariably are the same, a yellowish-green or light pale green; Y shows a greenish-brown or a darker shade of yellowish-green. Shades of brown and red are sometimes seen.

The index of refraction for γ is variable between the limits of 1.69 and 1.73. The smaller granules have, in general, the higher indices. The birefringence is also quite variable, ranging from 0.018 to 0.035, and the smaller crystals have a lower birefringence. The dispersion is inclined with $r > v$, and in most cases is stronger in the pigeonites of late crystallization.

The extinction angle $Z \wedge c$ is positive and varies from 21° to a maximum of 42° ; the lower angles were checked with a Fedorov universal stage. Crystals of the same size and in the same specimen usually have extinction angles agreeing within a few degrees of each other. However, within the same specimen, crystals of different sizes may vary as much as 20° . Figure 2 shows the irregularity of the extinction $Z \wedge c$.

The optic plane, as determined in large euhedral crystals, is parallel to (010). The optic sign is positive. The optic angle varies between the limits of 15° and 48° . There is a distinct break between the optic angle of pigeonite and augite. The grouping of the augite points in Fig. 2 is more regular than the pigeonite. It is evident that the only optical property which distinguishes pigeonite is the value of its optic angle. Pigeonites have a higher birefringence in general, but their average is that of augite. The extinction angle is generally lower than augite, but is not constant enough to be dependable.

Pigeonite contains inclusions of magnetite granules, plagioclase laths, shards and globules of glass, and empty bubbles and liquid filled holes. Pigeonite and magnetite grains occur intermingled in the reaction rims of hornblende and hypersthene. Augite and pigeonite occur as inclusions in plagioclase and augite.

Magnetite

Magnetite and ilmenite are present as inclusions in reaction rims, and in the groundmass of the Cascade andesites. Magnetites also occur as a product of weathering. Magnetite makes up about 3–10 per cent and

averages about 5 per cent of the volume of the rocks. Both magnetite and ilmenite are called "magnetite" below.

As inclusions, magnetite is most abundant in the plagioclases where it occurs without apparent arrangement. It occurs as isolated inclusions, or as inclusions in globules of crystallized groundmass, or in glass particles which in turn are included in the feldspars. Magnetite also occurs as an inclusion in the "unknown mineral," or oligoclase.

Magnetite is the most abundant inclusion in hypersthene. The particles are bunched together in clumps that look like large single grains. The magnetite grains are scattered without apparent system through the hypersthene host. Tiny dust-like magnetite is the only inclusion in the enstatite. A few inclusions of magnetite are found in the pyroxenes. The form is invariably dust-like, equant grains, which under the highest magnification appear as rounded grains.

Dust-like magnetite grains, masses of grouped small grains, true crystals, skeletal crystals, and equant grains of magnetite are found scattered through the glassy groundmass where the best formed and largest magnetite crystals are found. The largest are usually not over 0.5 mm. in diameter and many are too small to measure with the highest magnification. The most common form of magnetite is in equant grains; the square or rhombohedral outlines of octahedrons are not uncommon. "Iron crosses," isolated "feathers" of skeletal crystals, and "needles" as much as 0.5 mm. in lengths are commonly seen.

The magnetite is strictly opaque and in reflected light gives a grayish-black color. No good etch-figures are obtained with aqua regia.

Daly⁹ averaged 33 augite andesites and found 0.80 per cent titanium oxide (TiO_2) to be present. Almost every analysis of andesite in the petrographic literature shows titania to be present. Our analyses show from 0.53 to 1.10 per cent of titania. Portions of a number of specimens were crushed, sieved, and separated. Bromoform (specific gravity 2.8) was used to separate the feldspars, quartz, and glass from the amphibole, pyroxene, olivine, magnetite, and ilmenite. The magnetite was removed by the use of an electromagnet.

The remaining amphibole, pyroxene, olivine, and ilmenite gave a test for titanium. The pyroxene, amphibole, and olivine were separated and each tested. The microchemical tests on the pyroxene gave negative results. Six hornblende crystals were separated and individual tests were run. Titanium was definitely present in three crystals, but judging from the test, only in a small amount. In one crystal the tests were not positive. Two crystals gave negative results. Resorbed olivine crystals were crushed

⁹ Daly, R. A., *Igneous Rocks and the Depths of the Earth*, New York, 1933.

and the resorbed material separated by a magnet, but tested negatively for titanium.

Phenocrysts of plagioclase were crushed and the inclusions which looked like either magnetite or ilmenite were separated by hand under the microscope. The material was not attracted to the magnet, but gave a test for titanium. Portions of the glassy groundmass were picked out of the specimens and crushed. The magnet picked up all of the magnetite or ilmenite. Microchemical tests were made and titanium was found to be present, but apparently in lesser amounts than in the inclusions in the plagioclases.

As a result of these tests, ilmenite or titaniferous magnetite was found to be present in these rocks. The qualitative tests showed that there was more titanium present in the minerals of early crystallization, especially in the dust-like inclusions in the plagioclases.

Atmospheric weathering of the olivine, hornblende, and pyroxene in some specimens results in the production of secondary magnetite. It is dull black in color and lacks the luster and sharp outlines of the primary magnetite. It is attracted to the magnet and otherwise has the physical properties of primary magnetite with the exception of the crystal form.

Apatite

Apatite occurs only as inclusions in plagioclases and olivine.

Tridymite

Tridymite possibly occurs in the reaction rims of the hypersthene as small irregular, colorless, anisotropic masses. Small granules of tridymite are mingled with grains of magnetite in the reaction rims of the hornblende. Tridymite also has migrated to cavities in the rock.

Tridymite occurs in intimate association with the glassy base of the Cascade andesites. The tridymite occurs as tiny, colorless or faintly milky-white flakes. The flakes are not cleaved nor fractured, the edges are irregular but sharp and do not thin or wedge out. Tridymite makes up from 2 to 5 per cent of the volume of the rocks, the higher percentage occurring in aegirine augite-bearing rocks.

After crushing bits of the groundmass, it is possible to pick out tridymite fragments with a kerosene moistened hair. The index of refraction is between 1.47 and 1.477 which is too low for either quartz or cristobalite, and lower than the index of the glass in which the tridymite occurs. Though the optic angle cannot be measured accurately with a Johanssen's auxiliary lense and universal stage, it is estimated by the curvature of the isogyres to be between 30° and 40° , and is of positive sign.

The presence of tridymite in these andesites is not unexpected in view

of the observations on andesitic lavas elsewhere. No cristobalite was found except lining cavities.

Glass

Glass occurs as tiny shards and globules. It occurs included in hypersthene, augite, pigeonite, plagioclases, and as matrix in the groundmass.

In the feldspars the shards are sharp angled, irregular in shape and irregularly scattered, but the globules are of rarer occurrence. The glass inclusions are pale green or colorless. The included glass in turn contains inclusions of magnetite. Glass of the above type also occurs as inclusions in masses of groundmass which occur included in the plagioclases. Glass shards occur sparingly as inclusions in hypersthene, wherein they appear, in part, to have sometimes moved in along fractures. Light green colored glass has moved into and widened fractures in pigeonite. Glass occurs in the interspaces between the plagioclase laths and pyroxene granules in the groundmass.

The amount of glass in the Cascade andesites ranges from a few per cent up to 70 per cent; the average is about 16 per cent. Holocrystalline rocks are unknown, and all highly glassy rocks contain a notable percentage of crystals usually of plagioclase.

The colored glass varies from a dark gray through shades of green and brown to a very dark brown. The darker colors occur in rocks with a larger per cent of glass. Gray, pale green, and yellow-brown glasses are the most common. The gray color is the most common and is due to numerous small bubbles, none of which contain a liquid, and some are attached to grains of magnetite. In hand specimens the glasses may appear very dark but, when viewed under higher magnification, are found to be glasses of a light color, clouded with many grains of magnetite. The light green colored glass is most often found filling fractures and embayments in minerals; the glass of the groundmass is darker in color and between the two glasses is often a sharp dividing line.

The darker green, yellow, and brownish colors occur in glass in which there are but few grains of magnetite, whereas the lighter colored glass often abounds in magnetite grains.

The glass in these rocks has an index of refraction always greater than 1.482, which is too high for it to be tridymite. The dark green or brown glass has the highest index of refraction, ranging between 1.58 and 1.60. The index of the light colored glass ranged between 1.48 and 1.49.

Anemousite has an index ranging from 1.53 to 1.55 and might be mistaken for glass. By choosing increasingly darker colors, the range of indices from 1.48 to 1.55 and 1.58 to 1.6 are found. No glasses with indices between 1.55 and 1.58 are found. Barth¹⁰ (page 404) found values from

¹⁰ *Op. Cit.*

1.50 to 1.53 in some colorless basaltic glasses and states that this indicates that the glasses are essentially of silica, alumina, and alkalies. In the next paragraph he makes the statement that it is "not possible to demonstrate a definite relation between index and composition of the glass." Morey and Merwin¹¹ in their work on the relations between composition, density, and optical properties of glass come to the conclusion that from the properties of a glass—"no justification exists for an assertion that definite compounds exist in, and determine the properties of glasses."

Bubbles

Bubbles occur as inclusions in feldspar, pyroxene, and hypersthene crystals, and in the groundmass. Most of the bubbles contain no liquid. A few filled with a movable liquid occur in plagioclase and hypersthene crystals. Often little grains of magnetite are attached to bubble walls.

Groundmass

Groundmass fills embayments in enstatite crystals. Globules of groundmass are very commonly found in the feldspars, and in turn contain glass, crystallites, and microlites of plagioclase, and tiny grains of magnetite. Their appearance is the same as the surrounding groundmass. The crystallites in these globules of the inclosed groundmass have the same composition as those of the surrounding groundmass. Small globules of groundmass are sometimes present in the hypersthene.

Hematite

In the reaction rings around some olivine crystals a few deep red colored, tiny flakes are noted. Since the flakes were not produced by weathering, but by oxidation while the lava was cooling, and since goethite does not form above 130° C.^{11a} it is concluded that the stain is hematite.

Quartz, Opal, Calcite, Serpentine, and Chlorite

Quartz, opal, and calcite fill cavities and appear to have been products resulting from the final escape of gases and liquids during the cooling of the lava. Large, milk-white quartz veins and veins of calcite also occur. The calcite not infrequently lies on the quartz in the cavities.

The opal is found as small globules, usually fibrous with the fibers radiating from a common center. It is usually milk-white, though sometimes colorless. The opals are often loose in these cavities and can be picked

¹¹ Morey, G. W., and Merwin, H. E., *Jour. Opt. Am.*, **22**, 662 (1932).

^{11a} Tunell, G., and Posnjak, E., The stability relation of goethite and hematite: *Ec. Geology*, **26**, 333-343 (1931).

out with a pen knife. Also, weathering frees them and they are sometimes concentrated by streams, as at Opal Springs.

The serpentine and chlorite are deuteric; associated with ferromagnesian minerals or in lavas that are very fresh throughout. They occur in groundmass that is unaltered.

Secondary Minerals

Secondary weathering effects are comparatively rare in these rocks, being confined to the surfaces of the flows and along the major joint planes.

The soils derived from these rocks have a distinctive and characteristic red color. Rock exposures are green or red in color. The green is due to chloritic alteration of the ferromagnesian minerals, slightly more intense or of a deeper color on a weathered surface than upon a freshly broken surface.

The red-yellow colors are entirely weathered surface colors. They are due to iron hydroxide that has moved from the interior of the rock to the surface.

ROCK ANALYSES

	<i>Chemical</i>			
	O-92	H-173b	O-660	Ag 777
SiO ₂	53.27%	50.80%	62.92%	56.74%
TiO ₂	0.95	1.10	0.53	0.90
ZrO ₂	0.10	0.13	0.80	0.14
Al ₂ O ₃	18.00	20.93	17.79	19.14
FeO	4.69	5.41	3.49	3.63
Fe ₂ O ₃	4.35	3.90	3.02	3.93
MnO	0.11	0.12	0.06	0.11
NiO	0.05	0.005	0.007	0.005
MgO	6.81	5.01	2.41	4.61
CaO	8.53	9.24	4.93	6.86
SrO	0.00	0.00	0.00	0.00
BaO	0.01	0.00	0.00	0.007
K ₂ O	0.48	0.28	0.40	0.34
Na ₂ O	1.56	1.86	2.52	1.91
CO ₂	0.00	0.00	0.00	0.04
P ₂ O ₅	0.13	0.10	0.11	0.17
S	0.03	0.03	0.05	0.024
Cl	0.06	0.006	0.007	0.014
H ₂ O+	0.82	0.93	0.77	1.29
Totals	99.95	98.851	99.814	99.86

		<i>Norms</i>		
Quartz	13.2	9.45	31.147	22.39
Orthoclase	2.781	1.67	2.226	2.003
Plagioclases				
Albite	13.116	15.74	21.496	16.148
Anorthite	40.608	45.32	23.642	32.708
Pyroxenes				
Hypersthene	3.138	5.17	3.021	2.097
Enstatite	16.69	12.41	6.023	11.473
Diopside	.648	—	—	—
Corundum	—	.95	4.587	3.649
Magnetite	6.333	5.66	4.399	5.695
Ilmenite	1.803	2.09	1.007	1.702
Apatite	.358	.236	.28	.372
Zircon	.159	.20	1.191	.301
Calcite	—	—	—	—
Pyrite	.068	.065	.12	.048
Totals	98.902	98.961	99.139	98.586
Water	.82	.93	.77	1.29
	99.722	99.891	99.909	99.876
Anorthite	75.53	73.8716	52.44882	66.72432
Albite	24.39	25.6562	47.50616	33.04172

BYTOWNITE

LABRADORITE

Ware, Glen C. *analyst*.

HISTORY OF THE MINERALS

Origin

Hornblende was one of the first minerals to form, as shown by the presence of former hornblendes that have, in large part, been wholly destroyed or altered, indicating conditions in the lava unfavorable to their existence.

The conditions favorable to the formation of hornblendes are mineralizers, high pressure, and low temperature, all of which characterize deep seated magmas more than surface lava flows. Also the oriented arrangement of rock texture about them show the hornblendes were present while the lava was quite fluid. Movement within the fluid magma fractured the fully formed hornblendes. Figure 3.

A hydrolytic origin for hornblende appears established. Both hydrogen and fluorine must have been present in the magma in order that it should form. Allen, Wright, and Clement¹² established the necessity for the presence of water in a melt, in order that orthorhombic might change to

¹² Allen E. T., Wright, F. E., Clement, J. K., Minerals of the composition of $MgSiO_3$: *Am. Jour. Sci.*, 4th Series, **22**, 404 (1906).

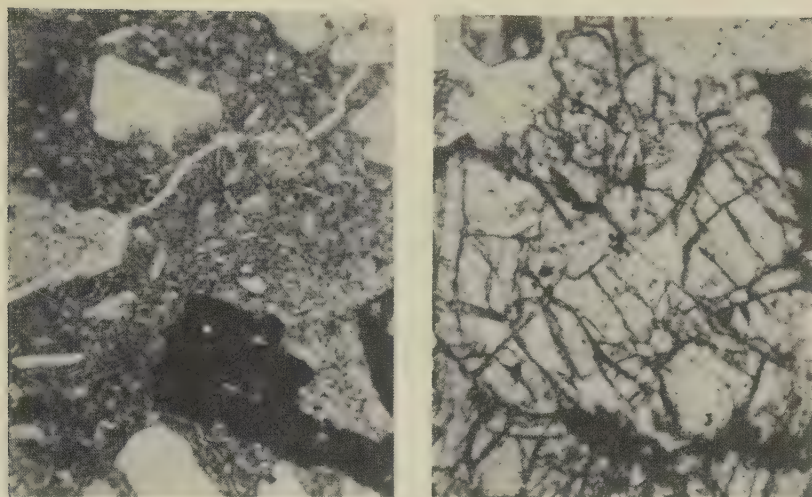


FIG. 3. (Left). Head of East Fork of Hood River, $\frac{1}{2}$ mile south of Mount Hood Lodge, S. line of Sec. 20, T 1 S., R. 10 E. Fractured, floated and much resorbed hornblende. Plagioclase zoned, center resorbed and replaced by groundmass. Upper part of field is a resorbed plagioclase and at top pigeonite and plagioclase are intergrown. $\times 110$.

FIG. 4. (Right). Fayalite. $\times 110$.

monocline amphibole. Such elements were only present in abundance before the magma vented onto the surface.

Emerging at the surface the gases were free to escape. The rate of loss was dependent upon the viscosity of the lava, being freer to escape in lavas of low viscosity, but more restricted in their circulation in lavas of high viscosity.

Not only was the hornblende formed under conditions of high pressure and in the presence of the original gases of the magmas, but it was formed at lower temperatures than those of the vented lavas. Allen, Wright, and Clement state that they found amphibole to change to monoclinic pyroxene with a continued rise in temperature which took place at an inversion point of 1150° . The reaction was not reversible. Shand¹³ states that deep magmas were never at a higher temperature than 1170° and in many instances were never higher than 870° . Lavas, upon reaching the surface, oxidize rapidly and increase in temperature. Jagger¹⁴ demonstrated at Hawaii that a change in temperature from 758° to 850° up to 1350° took place in a distance of from three feet below

¹³ Shand, S. J., *Eruptive Rocks*, 56 (1927).

¹⁴ Jagger, T. A., *Am. Jour. Sci.*, 214 (1917).

the basalt to the surface of the lava. Day and Allen,¹⁵ at Lassen Peak, California, found that the lowest temperature at which the andesites and dacites could be deformed was 850° in lavas that had lost most of their volatile material.

Not only were the hornblendes once present in these lavas in large numbers and under conditions that did not prevail at the surface, but those that have been preserved have been greatly altered. Kôzu, Yoshiki, and Kanu¹⁶ found that common hornblende heated to 750° changed to brown hornblende, due to loss of water and oxidation of the iron.

Winchell¹⁷ quotes Belovsky on heating hornblende and changing the extinction angle from 20° to 0.5°, and the color from green to dark brown. Both changes are probably due to oxidation FeO to Fe₂O₃. Weinschenk¹⁸ states that basaltic hornblende owes its color to ferric oxide and may contain up to 5 per cent titanium. He further states that green hornblendes turn brown when heated in air, and after heating the optical properties are the same as basaltic varieties. Graham¹⁹ heated green hornblende and found the extinction angle changed from 20° to 3°, and that the birefringence (Ng-Np) became much higher and the mineral acquired, after ten minutes heating, all the properties of basaltic hornblende. His green was found to contain 1.58 per cent, while the brown averaged only 0.46 per cent water. Graham (page 122) concluded that the change from brown to green was through the loss of water which changed the optical properties.

Allen and Clement²⁰ state that hornblende contains less water than the other amphiboles, and suggest that the water is not combined but dissolved. The conclusion is that the loss of hydrogen from the water in hornblende to furnish the oxygen to change ferrous to ferric iron is the change that takes place during the venting of the lavas.

Hornblende crystals from two specimens were taken and heated in open crucibles in an attempt to determine if the mineral in these rocks would behave in a similar manner to those cited above.

The specimens chosen were as near the average as possible, or of a light brown to a brownish-green color. The pieces were then placed in open crucibles and kept at a dull red heat. The results are given below:

¹⁵ Day, A. L., and Allen, E. T., Volcanic activity and hot springs of Lassen Peak, California: *Carnegie Inst. of Washington*, Publ. **360**, 51 (1925).

¹⁶ Kôzu, Yoshiki, and Kanu, Notes on the transformation of common hornblende: *Sc. Rep. Tohoku Imp. Un.*, Ser. III, **3**, 143-149 (1927).

¹⁷ *Op. cit.*

¹⁸ Weinschenk, Ernst; *Petrographic Methods*, 291-292 (1912).

¹⁹ Graham, W. A. P., Notes on hornblende: *Am. Mineral.*, **11**, No. 5, 121-123 (1926).

²⁰ Allen, E. T., and Clement, J. K., Role of water in tremolite and certain other minerals: *Am. Jour. Sci.*, 4th Series, **26**, 114 (1908).

Specimen	Hours of Heating	Z\C	Remarks
I	0	19°-17½°	Light brown pleochroism, distinct but not extreme.
II	0	20°-18½°	Brownish-green, pleochroism distinct but not extreme.
I	75	8°- 9°	Distinct brown color and pleochroism increased.
II	75	7°- 7½°	Light brown in color and all traces of green gone.
I	160	8°- 8½°	Black masses which look like magnetite. Both specimens the same in color and general appearance.
II	160	7½°	
I	317	5°- 6½°	Both dark brown in color and the black masses are increased in size. Outer edges of the crystals almost opaque.
II	317	5°- 5½°	
I	430	4°- 5½°	Outer portions opaque, inner portions a dark brown. Black masses much larger, especially in outer portions. Birefringence very high.
II	430	3°- 4½°	

The heating was stopped after 75, 160, 317, and 430 hours, respectively, and the material compared with the thin sections, and without exception, crystals identical in color, pleochroism, birefringence, and alteration were found. The appearance and characteristics were so similar that it was immediately apparent that the results produced in the crucible were comparable to those produced in nature.

The early crystallization of the crystals rich in fosterite and later crystallization of those crystals rich in fayalite is in agreement with their melting points. In any mix-crystal series, those having the highest melting points will crystallize first.²¹ The melting points of fosterite and fayalite have been determined by Bowen and Schairer.²² Natural fayalite from Rockport, Mass., melted at $1205 \pm 2^\circ\text{C}$. Fayalite from Ireland melted through a range of 10°C . from 1207° to 1217° . Fosterite was found to melt at 1890°C . Iddings²³ (page 85) gives the melting point of fosterite as greater than 1460° ; olivine from three different localities between 1395° and 1265° ; hyalosiderite between 1215° and 1220° ; hornblende between 1175° and 1180° ; and fayalite at 1055° .

The olivines also appear to have been present in the lavas prior to its issuance onto the surface. Their several melting points are in agreement with their order of crystallization, but imply higher temperatures than those indicated for the origin of hornblende. It is possible that much higher temperatures resulted from oxidation of the lavas in passing through the uppermost porous and water and gas-filled formations. Basaltic lavas (Coriba formation) several thousand feet thick underlie the Cascan formation almost everywhere in Oregon. The Cascan lavas may

²¹ Tyrell, G. W., *Principles of Petrology*, 74-75, London (1930).

²² Bowen, N. L., and Schairer, J. F., The system FeO-SiO_2 : *Am. Jour. Sci.*, **24**, 201-202 (1932).

²³ Iddings, J. P., *Igneous Rocks*, I (1909).

have melted these basalts on their ascent. Fenner²⁴ has found evidence that rhyolitic lavas flowing over the surface will melt and incorporate basaltic material which melts at 1000° or 1100°. Perhaps a magma, quite acid in character, bearing hornblende, melted its way through basalts and derived therefrom material for olivine, bytownite, and additional pyroxenes. Under such conditions, the hornblende may have been resorbed in large part and olivine, augite, and calcic plagioclases formed. The enstatite may be due to basaltic contamination of the emerging magma.

The hypersthene shows a magnesia-rich center and an iron-rich periphery. They definitely show that the composition of the magma was changing during their crystallization. According to Winchell²⁵ (page 177) the orthorhombic pyroxenes form a continuous insomorphous series between enstatite and hypersthene. The cores of our crystals show a composition at the beginning of crystallization of about 18 per cent FeSiO₃, and the outer portion, at the end of crystallization, a composition of about 45 per cent FeSiO₃. This enrichment of the magma in iron as crystallization proceeded is also shown by the olivine, magnetite, and the iron-rich glass.

The pyroxenes may have formed later, since they can be formed in a dry melt and at temperatures as high as 1150°. At depths any temperature above 550° would have been favorable for augite and unfavorable for hornblende. As the temperature increased upon emission of the lavas, it became more unfavorable for hornblende.

We may conclude that the hornblende was formed under temperatures below 850° at least, and perhaps below 550°; the augite, plagioclase, and olivine at temperatures at least above 550° and below 850°.

The final evidence on temperature is furnished by the tridymite. Tridymite indicates these lavas were at a temperature of at least 870°, and not greater than 1470°.²⁶

The evidence is that the Cascade andesites have changed from a deep-seated state of low temperatures, high pressures, and rising temperatures, to a surface state of rapidly rising temperature, falling pressures, and a rapid loss of volatile constituents.

RESORPTION

Hornblende

The weathered appearance of hornblende is deceptive because the sur-

²⁴ Fenner, Clarence N., A view of magmatic differentiation: *Jour. Geol.*, **45**, 158-168 (1937).

Tuffs and other volcanic deposits of Katmai and Yellowstone Park: *Trans. Am. Geophys. Union*, 18th Ann. Meeting, 236-239, *Nat. Res. Council*, Washington, D. C. (1937).

²⁵ *Op. Cit.*

²⁶ Wright, F. E., and Larsen, E. S., Quartz as a geologic thermometer: *Am. Jour. Sci.*, 4th series, **28**, 421 (1909).

rounding minerals are perfectly fresh. The hornblende has all been altered to the basaltic type, and is much absorbed. Zonal structure, if present, is destroyed by resorption.

The reaction products are rims of magnetite granules which may be pseudomorphic after the hornblende, pigeonite, microlites of feldspar, and tridymite. The prism faces are the most absorbed, the pyramid faces least, and the clinopinacoid and orthopinacoid are equally absorbed. The absorption has followed cleavages and fractures. Inclusions occur in the resorption products, but no inclusions are found in the unaltered hornblende.

In the ascent of the magma the hornblende moved through successive zones of lower pressures and higher temperatures to the surface. In them, the green hornblende might have undergone inversion to pyroxene, but no hornblende or pyroxenes have been found showing such inversion, neither does the hornblende found show absorption attributable to this stage.

In these rocks there is found just a reverse of Bowman's reaction series,²⁷ which however, is applicable only to intrusive magmas where fractional crystallization takes place with an increasing amount of volatile constituents. The hornblende was formed early and was carried upward rapidly to the surface.

Certainly conditions became unfavorable for hornblende long before it reached the surface and much of it, but not all, was lost enroute. Upon arriving at the surface it was changed first to brown or basaltic hornblende. The brown hornblende secondly was not in equilibrium in the magma and it underwent rapid absorption. The absorption was so vigorous that many of the hornblendes were totally dissolved, others remained as areas with only reaction rings of magnetite, and others were preserved in an advanced stage of demolition. Even so, hornblende is only preserved in those lavas that chilled quickly as in thin, porous trachytic lavas. In all others it disappeared altogether.

The absorption of basaltic hornblende took place when the lavas were stationary, as shown by the manner in which the magnetite grains hug the absorbed crystals and so preserve the form of wholly dissolved hornblende.

Olivine

Olivine, like hornblende, shows absorption varying from complete solution to partial absorption. The absorption affected mostly the larger grains, and few of the tiny grains in the groundness. The absorption is

²⁷ Bowman, N. L., Ore deposits of the western States: *Am. Inst. Met. & Min.*, 110, (1933).

proportional to the surface area of olivine exposed to solution. Absorption began either before the lavas vented or immediately thereafter. In each rock one type of resorption predominates, but varies with different rocks. The minerals produced by absorption are principally pigeonite, much magnetite, and some hematite, iddingsite, and bowlingite.

Since much of the pyroxene in these rocks is pigeonite, and most of the pigeonite has resulted from hornblende and olivine, it appears as though the olivine was once quite abundant in these rocks. The olivines remaining are, like hornblende, only residual remnants.

Fayalite

The very small grains of fayalite, Fig. 4, unlike the other small grains of olivine, have suffered little or no absorption. The glass indicates that iron and silica were in solution in the liquid magma even to the ultimate final freezing of the rock. Because fayalite contained no elements not always available in the unused melt, and because it is a saturated compound, it suffered no reaction. Bowen and Schairer²⁸ found it is stable in the presence of quartz; thus, explaining why olivine found in siliceous rocks is always fayalite.

Magnetite

Magnetite was formed by crystallization directly out of the melt and as a by-product of reaction, being one of the main products in the breakdown of the olivines. Thus as silica was added to the olivine in a melt rich in iron to form pigeonite, the unneeded magnetite was left behind.

Equant grains of sooty clumps of magnetite occur in many reaction rims. It is the most abundant of the released minerals occurring in amounts that almost hide the tiny grains of pigeonite and feldspar. Magnetite forms rings or zones associated with small granules of quartz around augite. Magnetite forms as a by-product of other forms of reaction as in the production of pigeonite.

Hornblende when absorbed along the borders or within cleavages and fractures is replaced by sooty, black aggregates of magnetite granules, and in most cases magnetite entirely replaces the hornblende. In plagioclases which have had the periphery absorbed and then replaced by new feldspar, the magnetite occurs in larger amounts than in the core.

When pigeonite is formed from olivine, magnetite is rarely formed, or, if it does form, it occurs as a very narrow surrounding ring. Absorption of olivine without the formation of pigeonite results in much magnetite being formed as a halo about the residual core, often preserving the form

²⁸ *Op. Cit.*

of the domes and pinacoids. The tiny magnetite grains appear with beginning absorption and increase until they form a dense packed mass entirely around an extensively altered crystal.

Pigeonite

Though the emerging lavas carried crystals of olivine and bytownite, yet they also were rich in silica. The silica was, perhaps, that of the original magma in which the hornblende formed. Not all of the silica was used in reaction with the incorporated basalts obtained by melting an outlet to the surface. The silica, therefore, moved onwards towards the completion of its equilibrium with the incorporated minerals.

Bowen and Schairer²⁹ (page 203) state that in "the presence of the volatile constituent water, ferrous oxide and silica, under proper conditions, combine in a ratio that approaches, though it does not attain, the metasilicate ratio."

The principal product of reaction of olivine and hornblende with the magma is pigeonite. Very little augite occurs in these rocks and since pigeonite can only occur by the reaction of a siliceous melt on pyriboles and olivine, it is possible that there was once much olivine present in the magmas of these rocks.

"In a system containing the components of an olivine and pyroxene-bearing rock, there will always be a reaction field, the melts of which solidify in such a way that the olivine is resorbed during the cooling."³⁰

Masses of pigeonite occur as the reaction products of olivine, hypersthene, and augite. In some of these it is clear that the pigeonite is the result of reaction with the enclosed mineral, as shown by resorbed boundaries; in other cases there is no evidence of reaction except the presence of the pigeonite. Some pigeonite grains occur isolated and may have crystallized directly from the magma.

Augite is often partially resorbed and replaced by pigeonite even to the extent of forming armored shells of closely packed granules, partially or wholly around the augite.

The common reaction on olivine grains, especially fosterite, produces pigeonite (Fig. 5). Fayalite grains do not show this reaction at all. The descriptive term "armored relics" as first used by Eskola and then by Barth³¹ (pages 381-382), in his description of the Pacific lavas is highly applicable to Cascan lavas. When pigeonite completely enclosed an olivine crystal, resorption was halted. Crystals of olivine that were only

²⁹ *Op. Cit.*

³⁰ Anderson, Olaf, The system of anorthite, fosterite, silica: *Am. Jour. Sci.* 4th series, 39, 407-454 (1915).

³¹ *Op. Cit.*

partially surrounded by pigeonite were not so protected and resorption continued, but in many cases the rock solidified before the reaction was complete.

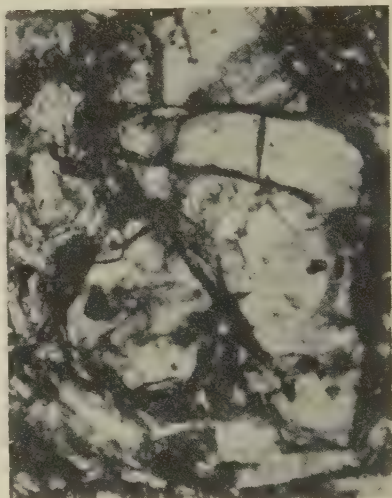
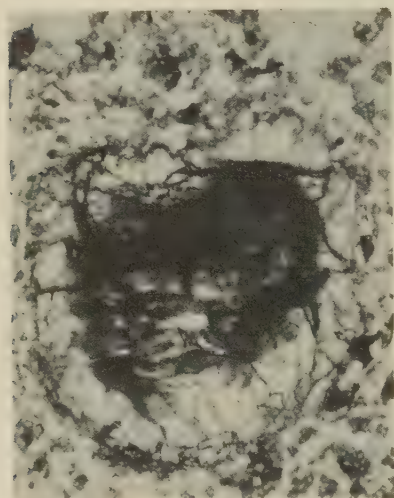


FIG. 5. (Left). South of Mount Hood, near Government Camp on Multorpor at an elevation of 4675 ft. and from a flow 25 feet thick. Olivine surrounded by an armour of pigeonite granules. Pigeonite crystallites are interspersed among labradorite laths. $\times 110$.

FIG. 6. (Right). N.E. $\frac{1}{4}$ Sec. 15, T 6 S., R 7 E. elevation 3450 ft. Olivine is deeply embayed and replaced in embayment and at the edges by pigeonite. $\times 110$.

The pigeonite which surrounds the olivine crystals has the same optical properties as that which is not closely associated with the olivine. Barth³² found a crystallographic relationship between the olivine and the pigeonite; the *c*-axis of the pigeonite and the *c*-axis of the olivine being mutually perpendicular. In our rocks the pigeonite forms along or around the crystal faces of the olivine crystals with apparently no preference, for the composition plane may be either a pinacoid, dome, or base. In every case, however, the contact is sharp and distinct with no other intervening minerals (magnetite) between the crystals (Figs. 5 and 6). Barth also found that the composition plane was usually the base of the olivine crystal. This is not the habit in these rocks.

Therefore, the evident explanation of the pigeonite was the increase of silica in the residual magma to such amount that it began to react with the olivine. Where thin rings of magnetite surround an olivine and this ring separates the olivines from the pigeonite, we see evidence of a movement towards the formation of a pyroxene and feldspars, and then a sud-

³² *Op. Cit.*

den over-saturation by silica. In other cases we find the pigeonite joins the affected olivine with a sharp contact showing that as fast as silica became available it altered the olivine to pigeonite. The chemical composition of pigeonite averages 50.46 per cent SiO_2 .³³ The SiO_2 content of olivine as given by Iddings³⁴ is not higher than 42.3 per cent. This shows the amount of enrichment of silica necessary to alter olivine to pigeonite. The pigeonite was formed of fosterite-rich olivine because such olivines being formed first, were in existence longer than the fayalite, and therefore available victims of attack before the fayalite was formed.

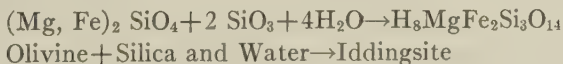
Bowlingite and Iddingsite

Ross³⁵ states that iddingsite is a deuterite mineral derived during the final cooling of the lava in which it occurs by a reaction between gases, water, and olivine, under oxidizing conditions. It is always formed after a magma comes to rest because iddingsite, though very brittle, is never fractured nor distorted by flowage. Ross suggests that the water was sealed in by freezing of the lava surface.

Barth found iddingsite as a deuterite mineral in some of the Pacific lavas. He states that iddingsite was noted in the groundmass in places where olivine probably never existed. In the Cascade andesites, iddingsite is found only in reaction rims around olivine.

Aurousseau³⁶ explained these minerals as a result of an oxidation by volatiles during eruption. Campbell and Stenhouse³⁷ found that even in the "freshest specimens" iddingsite after olivine is found, but offered no explanation.

Both of these minerals occur in reaction rims of olivine, in our rocks, that show no sign of weathering, consequently, the water contributed to their hydrous nature must have come from the rock itself. Since both minerals are metasilicates, we conclude that where water and silica became available these minerals were formed instead of pigeonite by the following reactions:³⁸



³³ *Op. Cit.* Average of three analyses.

³⁴ Iddings, J. P., *Rock Minerals*, 380, New York (1911).

³⁵ Ross, C. S., and Shannon, E. V., The origin, occurrence, composition and physical properties of the mineral iddingsite: *Proc. U. S. National Mus.*, **67**, 1-19 (1925).

³⁶ Aurousseau, M., Analysis of three Australian rocks: *Proc. Linnean Soc. New So. Wales*, **51**, part 4, 617-618 (1926).

³⁷ Campbell, R., and Stenhouse, A. G., Occurrence of nepheline and fayalite in the phonolitic trachyte of the Bass Rock: *Trans. Edin. Geol. Soc.*, **8**, part 1, 128-129 (1932).

³⁸ *Op. Cit.* This formula is given as the questionable composition of iddingsite. The composition of iddingsite given by Larsen (*U.S.G.S., Bull.* **846**, 131) is the same but in a different form: $\text{MgO} \cdot \text{Fe}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot 4\text{H}_2\text{O}$. Larson does not question this composition.

The "basaltic hornblende," the world-wide experience with volcanic eruptions, and the vesicular character of the lavas, indicate much water was available to form iddingsite and bowlingite. The fact that only a few olivines reacted to form iddingsite and bowlingite indicates that much of the water immediately escaped from the lavas at the time of their eruption. This conclusion seems to be in agreement with Bowen's³⁹ statement that water does not enter into crystallization at high temperatures, and that the reaction between the olivines and the water took place just before the issuance of the lavas at the surface.

Enstatite

Enstatite rarely occurs and then only in large crystals. It contains a few grains of magnetite and has been largely absorbed to form hypersthene. Some of the enstatites have embayments filled with groundmass. Orthorhombic pyroxene was shown by Allen⁴⁰ (pages 385, 414-415), to be unstable at high temperatures, and that enstatite changes to the monoclinic pyroxene above 1260°. Clark⁴¹ says: "The presence of enstatite in an igneous rock is evidence that the final crystallization took place at relatively lower temperatures . . ." and that the effect of iron on modifying the properties of these silicates "is undetermined."

Hypersthene

The hypersthene has been greatly resorbed. The reaction effects are so abundant and of such a character as to suggest that it was once more abundant in the lavas, and its absence in certain flows is wholly the result of resorption (Fig. 7).

Armored relics of hypersthene wholly and partially surrounded by augite and pigeonite are common. Some have embayments entirely filled by pigeonite grains admixed with a few plagioclase laths. In cases where a single augite or pigeonite grain formed from a hypersthene crystal, the contact between the two is sharp with no intercollated material. Magnetite grains occur around hypersthene boundaries, but less abundantly than around hornblende and olivine. In cases where complete absorption has taken place, magnetite-soot forms a pseudomorph after hypersthene.

Augite

A few large augite crystals occur isolated from olivine, hornblende, and orthorhombic pyroxenes, and are surrounded by reaction products.

³⁹ Bowen, N. L., *Ore Deposits of the Western States*, 112-113, New York (1933).

⁴⁰ *Op. Cit.*

⁴¹ Clark, F. W., *Data of Geochemistry, U.S.G.S., Bull. 77*, 380 (1924).

Such crystals may have formed early in the magma and perhaps represent resorption products of the basalts. Other augite crystals are deuteritic products. Reaction rings are common, but embayments are rarely seen.

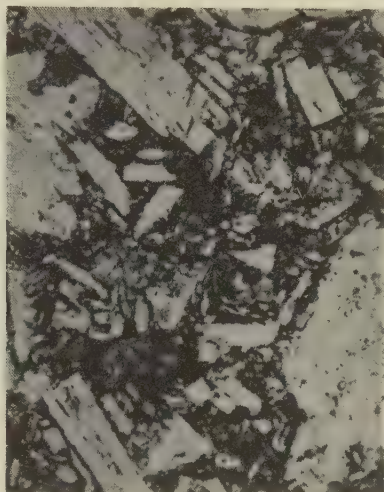
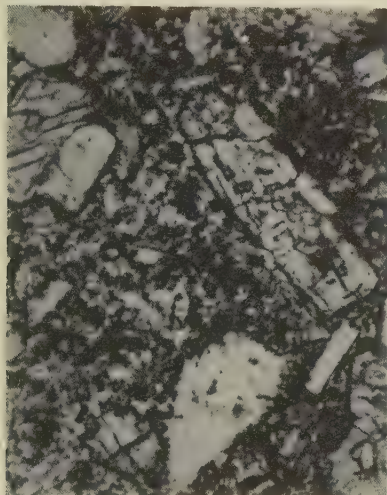


FIG. 7. (Left). Sugar Loaf Mountain, West of Mount Hood, Sec. 14, T 2 S., R 7 E., elevation 3500 ft. Lower center a resorbed labradorite, upper right hypersthene embayed by feldspar and replaced by pigeonite along fractures. Upper left augite. $\times 110$.

FIG. 8. (Right). East of Alkalie Butte, one half mile from Trout Lake on Mill Creek Trail. T 9 S., R 8 $\frac{1}{2}$ E. Bytownite phenocrysts in groundmass of labradorite laths and pigeonite grains. $\times 110$.

The principal resorption product of augite is pigeonite. Such pigeonite forms closely packed granules around the partially dissolved augite crystals. Intermingled with the pigeonite granules are small grains of magnetite and tridymite. The amount of absorption of the augite is much less than that for any of the minerals discussed above.

The release of mineralizers by the dissolution of the hornblende may have aided in the early formation of the augites, perhaps before the lavas emerged onto the surface.

The above suppositions seem pertinent in order to explain why augite formed and remained relatively stable along with the more stable pigeonite. In view of the direct formation of pigeonite from most of the minerals, including augite, it is not reasonable to suppose that augite is an intermediate product. It appears more reasonable to suppose that augite formed earlier and under low temperatures and higher pressures, and

became itself unstable later at the higher temperature and lower pressure that prevailed conjointly for a time in the lavas.

	Analysis of Augite ^a and Pigeonite ^b	
	Augite	Pigeonite
FeO, Fe ₂ O ₃	6.79	11.84
CaO	22.07	9.24
Na ₂ O	0.87	0.23
SiO ₂	48.76	50.46

^a Averages taken from Iddings, J. P., *Rock Minerals*; Clarke, F. W., *Data of Geochemistry*; and Clarke, F. W., *U.S.G.S. Bull.* 591.

^b Averages taken from Winchell, A. N., *Elements of Optical Mineralogy*, Pt. II.

Augite is lower in iron and silica, and richer in soda and lime than pigeonite. When resorption took place the latter went into plagioclases and the deficiency of iron and silica was supplied by the magma to form pigeonite.

An attempt was made to determine if the plagioclase crystals near the resorbed pyroxenes were of different composition than those crystals which had formed at a greater distance. Though the plagioclases showed a slight variation, the results were not conclusive.

Pigeonite

The pigeonite itself is almost always unaltered. Deep embayments such as are found in the other pyroxenes are not found in crystals of pigeonite. The larger crystals sometimes have narrow reaction rings with magnetite as the chief resorption product. Occasionally small granules of quartz are found between the magnetite grains.

Plagioclase

Resorption has affected the larger and more calcic crystals; the sodic crystals occasionally show a slight solution of their periphery, but never any embayments.

The residual "juices" dissolved passages through the outer zone of the more sodic shell, through which it entered and reached the calcic interior of the crystal. Three crystals were found where the solution entered a passage and removed the core of the crystal, filled it by material of the same composition as the groundmass, and left the outer zone as a shell. Inclusions of olivine, fayalite, augite, and pigeonite were resorbed and removed so that the sodic plagioclase added does not contain the minerals found in the outer zone.

In other crystals after the core was formed it was embayed and the embayment filled by the same material that formed the outer zone. The

normal zoning of calcic core, grading outward to sodic borders, is explained as the result of progressive crystallization with reaction. No oscillatory zoning is found; hence, the magma moved steadily upwards to the surface.

Tridymite

Tridymite is found in the reaction rims, the groundmass, and in cavities in the rocks. The tridymite in the above occurrences indicate that it formed when reaction set in on all of the minerals and continued to form until the lavas finally froze. Its presence in the cavities indicates its movement along with the escaping mineralizers.

Bowen⁴² has shown that an incomplete reaction between olivine and silica tends to enrich the liquid in silica. The final products are olivine, pyroxene, and quartz. Bowen called this quartz a "released mineral." Some of the quartz in our rocks may be of this origin; however, the presence of so much tridymite and the chemical analyses indicate an excess of silica. The silica is that which under less rapid cooling would have converted more of the orthosilicates to a higher state of silica saturation, a reaction that was halted by the solidification of the lavas.

The tridymite is proof of the type of process, whereby original olivine and other minerals were resorbed to form pigeonite: a process whereby tridymite occurs with olivine and in rocks analytically rich in silica, and wherein the olivine is much absorbed. The olivine should not be weighed in favor of a basalt classification, but in favor of an andesite classification for the Cascan rocks.

ORDER OF CRYSTALLIZATION

Hornblende

Evidence has been given to show that hornblende was present in the magma at an early stage, that it changed to the basaltic type immediately after extrusion and then suffered almost complete absorption.

The early crystallization of hornblende is shown by its presence only as phenocrysts. Many of them are fractured and surrounded by flowage structure. Some plagioclase laths occur as inclusions in hornblende.

Hornblende by reaction gave rise to magnetite, pigeonite, tridymite, and perhaps plagioclase, but these reactions took place at a late stage when the lava was stagnant.

Olivine

Olivine must, also, have been fully formed before pigeonite started to

⁴² Bowen, N. L., The system of MgO-SiO₂: *Jour. Geol.*, **30**, 182 (1922).

form, except for some small olivine crystals which may have resulted by recrystallization of formerly dissolved olivine during later cooling.⁴³

Olivine formed before extrusion and continued to form for a long time thereafter. However, even the smallest olivines were in existence when the calcic plagioclases formed, and in the latter only the smaller olivines are rarely found as inclusions. These inclusions are always included in the outer sodic rims of the larger crystals.

Olivine, by reaction, gave rise first to magnetite, augite, and feldspars probably before emission of the lavas. Later and probably after emission, the olivine yielded pigeonite, iddingsite, and bowlingite by reaction.

Fayalite

Fayalite apparently formed after the other olivine had largely formed. The conditions were favorable for fayalite to form in the magma before its eruption and it might have so formed along with the early magnetite. However, the small grains of olivine and fayalite occur only as inclusions in the larger plagioclases, near their outer borders.

Augite

Some of the large augite crystals were formed before emission of these lavas and apparently at about the same time as the calcic plagioclases. It was not derived by direct inversion of hornblende, though some may have formed by recrystallization of wholly dissolved hornblendes. The evidence of its early crystallization is its presence in large phenocrystic crystals. Some crystals of augite were in existence when the sodic rims of the plagioclases formed.

Augite does not occur abundantly and none of it was stable, giving way by reaction to magnetite, tridymite, and chiefly to pigeonite which formed at a late stage after the lava had erupted.

Enstatite

Enstatite crystallized as early as the hornblende and certainly earlier than the olivine and augite. This is shown by large crystals and the absence (except magnetite) of inclusions within the smaller grains. It formed early probably in lieu of the formation of diopside.⁴⁴ It was quite unstable and is mostly resorbed to form hypersthene.

⁴³ Bowen, N. L., and Anderson, O., The binary system MgO-SiO_2 : *Am. Jour. Sci.*, **37**, 487 (1914).

⁴⁴ Barth, Tom F. W., Crystallization of pyroxenes from basalts: *Am. Mineral.*, **16**, No. 5, 195-208 (1931).

Hypersthene

In iron-rich rocks hypersthene is found more commonly than enstatite. The large number of inclusions in the hypersthene as compared to enstatite and olivine indicates that it formed at a later time. Hypersthene formed shortly before the emission of the lavas, because it occurs as phenocrysts and is aligned with the flow structure of the rock. It formed contemporaneously with many of the plagioclases because it is molded on them, especially in the glassy rocks. On the other hand, laths of plagioclase are oriented around the hypersthene.

Small globules of groundmass are present as inclusions in the hypersthene which appear to have no connection with the surrounding groundmass. Such included groundmass contains, like the matrix groundmass, crystallites of plagioclase and magnetite dust. This material was apparently liquid included in the crystallizing hypersthene which later crystallized. Hypersthene was quite unstable and is mostly resorbed, except in the very glassy rocks.

Pigeonite

Pigeonite definitely formed after the emission of the lavas as a reaction product of hornblende, olivine, enstatite, augite, and hypersthene. It formed to such an extent that it almost entirely replaced its parental minerals. Where hypersthene and olivine remain it often jackets them completely. Elsewhere it occurs in clusters over an area once occupied by the primary minerals and elsewhere it is scattered throughout the lava. It continued to form to the very end, and would ultimately have eliminated all the above minerals had the lava not frozen. It is found mingled with the smallest plagioclase laths. Larger pigeonite grains include magnetite, plagioclase, glass, groundmass, and bubbles. The large abundance of inclusions prove it was one of the last minerals to crystallize. It occurs only in the outer sodic zones of plagioclase.

Plagioclase

The calcic plagioclase started to crystallize before emission of the lavas as shown by its presence in flows almost devoid of any other mineral, wherein it forms mesh-work or jack-straws-like arrangements from amid which all the residual liquor is drained. In most of the glassy rocks, plagioclase occurs as phenocrysts. No inclusions except bubbles or tiny magnetites occur in the calcic cores of the plagioclases, but in the outer sodic zones are a few inclusions of olivine, fayalite, augite, pigeonite, magnetite, glass and groundmass. Plagioclase continued to form to the very end, but as more and more sodic varieties. Tiny microlites occur in the groundmass and some glassy areas appear to be oligoclase on the verge of crystallizing (Fig. 8).

Magnetite

Magnetite started to crystallize at a very early stage and continued to form either by direct crystallization out of the iron rich lava, or as reaction products. It occurs as inclusions in all minerals. All of the iron was freed from solution in all but a few of the lavas. Every gradation from green to brown glasses to groundmass with crystallites of magnetite to good crystals of magnetite included or excluded in other minerals are found.

The presence of iron probably accounts for the differences in the colors of the glass. When there are large numbers of magnetite grains or crystals present, the glass is usually one of the lighter colors, or even colorless. When there are but few magnetite grains present the glass is a darker shade of green or brown due to the state of oxidation of the iron.

Tridymite

Tridymite occurs in the reaction rims of many minerals and in the groundmass and in cavities in the lava. It was one of the last minerals to crystallize.

Glass



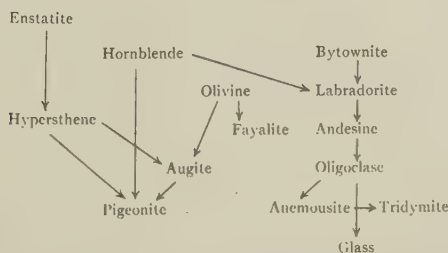
FIG. 9. Hill in Sec. 25, T 3 N., R. 6 E. at elevation 3710 ft. Large pigeonite grains above and below; between plagioclase. The groundmass of glass is black from magnetite crystallites. $\times 110$.

Glass containing bubbles was the last substance to solidify. The amount of glass varies inversely with the amount of feldspar, and the largest feldspars, though fewest in number, occur in the glassy rocks. The color of the glass has been referred to above.

Summary

The time of formation of the various minerals is shown in the following table. It is clear that these rocks did not follow the normal order of

Stages	Magmatic Chamber	Conduit	Vent	Erupted Lava	High Temperature Surface Lava	Cooling Lava	Freezing Lava
Hornblende	_____	_____					
Ilmenite		_____	_____	_____			
Apatite	_____						
Bytownite		_____	_____	_____			
Labradorite		_____	_____	_____			
Olivine		_____	_____	_____			
Fayalite		_____	_____	_____			
Andesine				_____	_____		
Oligoclase					_____	_____	_____
Anemousite							_____
Enstatite	_____	_____					
Hypersthene		_____	_____	_____			
Augite		_____	_____	_____			
Pigeonite			_____	_____		_____	
Magnetite					_____	_____	
Glass							_____
Tridymite					_____	_____	



crystallization as given by Bowen,⁴⁶ except for the plagioclases. Here hornblende and enstatite preceded the olivine.

⁴⁶ *Op. Cit.*

CONCLUSION

The rocks of the Cascan formation are made up of lavas with a diverse mineral composition. The minerals such as olivine, hypersthene, enstatite hornblende, plagioclases, and quartz occur in combinations that ordinarily define many different kinds of rock. It is true that at one place the mineral components of a certain lava warrant such names as olivine basalt, quartz basalt, enstatite andesite, andesite, dacite, and other names. Yet if these lavas are studied as a whole it becomes plainly evident that each was in the process of attaining equilibrium and was frozen at some place along this route. For example, the calcic plagioclases were moved rapidly toward a much more sodic type; the olivine, enstatite and hornblende were all unstable and moving toward pigeonite; the silica in the rock was rapidly transforming the orthosilicates to metasilicates; and the low temperature hornblendes were going over into high temperature pyroxenes, principally pigeonite. Detailed study of these rocks makes it difficult for the student to be satisfied with viewing a single lava flow occurring in a large assemblage of flows as an independent rock species. He sees it as only a transition stage in a process moving towards a goal.

Rocks not considered in the study are lavas of the late Pliocene or Pleistocene age that occur on the slopes of the younger volcanoes of the Cascade Range or as intracanyon flows. Such lavas differ radically from those of the Cascan formations as described by Hodge elsewhere.

Magmatic differentiation was far advanced before the magmas vented to the surface. Enroute they were affected by the walls of the conduit and on the surface experienced a profound change in pressure and temperature. Consequently, the usual reaction series as found in rocks which cooled slowly under great load and at great depths are not found.

The rocks were apparently derived from a magma that was rich in silica, soda, and alumina; it had a composition analogous to dacite. This magma approaching the surface passed through pyroclastics and some lavas that contained entrapped water and air with which it reacted by oxidation, resulting in a large increase in temperature. As a consequence, it was able to dissolve its way towards the surface. Basalts of the Coriba formation underlie much of the Cascan formation, and in consequence the dacitic magma was made more basic by the basaltic material. The passage through the basalts, however, was so rapid that equilibrium was not immediately attained. Upon the surface, a rapid increase in temperature aided the attainment of equilibrium, but the equally rapid sequential chilling abruptly interrupted the process.

Consequently, we find lavas that followed opened vents and retain their original dacitic character; elsewhere lavas that rose so rapidly that the hornblende was not resorbed; other lavas in which the olivine,

bytownite, and enstatite of basaltic derivation had not reacted with the dacitic magma before consolidation; and still other lavas in various stages of more completed reaction.

The calcium enrichment of the silicic ascending magmas indicated the ultimate formation of pigeonite in lieu of olivine, hornblende, enstatite, augite, and hypersthene, limited only by the available magnesia, and of plagioclase. The result would have been a pigeonite-plagioclase rock.

The phenocrysts of calcic plagioclase, olivine, enstatite, augite, and magnetite is the best evidence of pre-extrusion crystallization.

The plagioclases are not zoned, but do show a progressive change from center to periphery. This means that stagnation existed for a period prior to eruption so that equilibrium of the plagioclase in the magma took place. Upon arrival at the surface, the change towards equilibrium was more vigorous as shown by the resorption by embayments of the calcic cores, even to the extent of breaking through an outer rim of sodic plagioclase.

The lack of oscillatory zoning in the plagioclase shows the lava ascended steadily upwards, too fast for crystals to sink into lower zones. The first plagioclases to form were bytownites, which formed soon after the solution of the basalts. Thereafter, the soda of the dacites reacted rapidly with the bytownite. Consequently, there is a gradation from bytownite to oligoclase in the lavas.

The presence of bytownite and olivine defines an olivine basalt. The rock, however, is one not in equilibrium, and if it were, its feldspar would be labradorite.

The lavas are low in ferromagnesian minerals. The plagioclases average labradorite. Where reaction is complete, the lavas are labradorite (average) and pigeonite, a basic andesite.

In summary it was found that every specimen of these rocks had its own individual range and amounts of the different plagioclases. Also every large crystal has its own characteristic zones of various compositions and varying replacements. The rocks are alike as here described, but vary so much that one hesitates to lay down a definition for them. Attempts were made to group these rocks on the basis of the feldspars, and the largest group contained four specimens.

These rocks do not follow the normal reaction series and no gradational settling is evident in any of these flows.

The character of the lavas change not only from flow to flow, but within a single flow. Near the base more complete reaction has taken place, and particularly in flows 200 feet thick. The analyses given in the paper are from holocrystalline rocks selected from near the base of flows.

DEHYDRATION OF POLLUCITE

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ABSTRACT

Samples of pollucite from two localities have been studied under the microscope and by means of x-rays before and after dehydration. This study is further evidence that the water present in the mineral is not an essential part of the crystal lattice. The shape of the dehydration curve is not, in this case at least, a reliable means of deciding the role of water in a mineral.

Pollucite still remains, nearly a hundred years after its discovery by Breithaupt in 1846, the only mineral which contains the rare alkali cesium as an essential constituent, the Cs_2O content being as high as 36 per cent. Although pollucite has been considered to be of very rare occurrence, there is good reason to believe that it will be found at many more localities than the ten from which it has so far been reported: Hebron, Rumford, Buckfield, Greenwood, Mt. Mica and Newry in Maine; Leominster, Mass.; Tin Mountain, South Dakota; San Piero on the isle of Elba; and Varuträsk, Sweden. Massive pollucite is extremely difficult to distinguish from quartz in the field. It has no cleavage and has nearly the same hardness and luster as quartz. Under the microscope, however, it is readily distinguished, being isotropic. It is of considerable interest to the geochemist that at all the localities listed above, pollucite occurs in pegmatites rich in lithium minerals, although pollucite itself contains very little lithium. In other lithium-bearing pegmatites, its presence has probably been overlooked.**

Pollucite was long thought to be of fixed composition, but recently Richmond and Gonyer (5) showed that the composition is variable, and suggested the formula $(\text{Cs}, \text{Na})_{14+x}\text{Al}_{14+x}\text{Si}_{34-x}\text{O}_{96} \cdot n \text{H}_2\text{O}$, with $x=0, 1$, or 2, and with n variable between 4 and 9. The observed water content, which varies from 1.5 to 3.8 per cent, increases as sodium replaces cesium in the mineral. The dehydration of pollucite was studied in 1936 by Strunz (6), who considered that the dehydration curve (reproduced in Fig. 1) obtained by him on material from Elba was that of a typical zeolite, indicating that the water present (2.6 per cent in his sample) was not bound in the crystal lattice. He confirmed this by the observation that an x-ray powder photograph of material dehydrated at 500° was identical with that taken on the original sample. He found, how-

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** Dr. W. T. Schaller informs us that pollucite, in very small amount, has been identified from Pala and Mesa Grande, California. Here, as in the other pegmatites, it is associated with beryl containing cesium and with lithium minerals.

ever, that the mineral, unlike typical zeolites, did not, after dehydration, take up water on exposure to air saturated with water vapor.

The dehydration of pollucite seemed worthy of further study, since Strunz's results are in disagreement with observations by Wells (7), who reported that material from Hebron lost only 0.01 to 0.03 per cent at 170°. We have carried out dehydration experiments on two samples of pollucite, one from Hebron, Maine, and one from Elba, and have studied the product of dehydration under the microscope and by *x*-rays. The sample from Hebron (No. 2677 of the Brush Collection of Yale University) was part of the material analyzed by Wells. It was fresh and

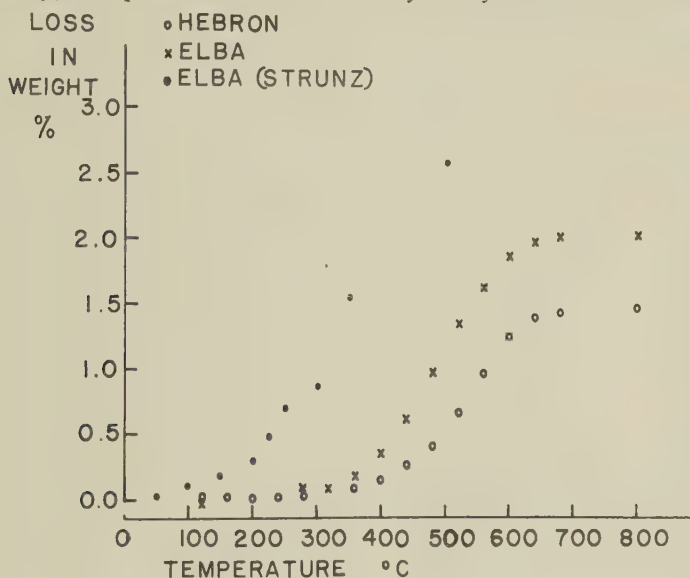


FIG. 1. Dehydration curve of pollucite.

transparent and of ideal purity. The sample from Elba (No. R2952 of the U. S. National Museum) was also transparent but contained a small amount of very fine grained material present along cracks. The sample was easily purified by floating off the foreign material in bromoform, and was then dried at 110°.

In the dehydration experiments, a weighed amount of the powdered mineral was placed in a platinum crucible and heated in an electric furnace, the temperature being held constant to $\pm 5^\circ$. When the loss in weight was less than 0.02 per cent over a 24-hour period, the temperature was raised.

Our results are given in Table 1 and Fig. 1, Strunz's results being also given for comparison. Both samples we studied showed practically no

loss in weight below 300° and were completely dehydrated at 640°. In the temperature range 300–640°, most of the loss in weight at each temperature occurred in the first twenty-four hours, but the sample continued to lose weight slowly. It seems probable that pollucite could be dehydrated completely at 450° or lower, but that a long time would be required. We cannot account for the difference between our results and those of Strunz, especially since he used a very short period of heating at each temperature.

TABLE 1. Dehydration of pollucite

Temp. °C.	Total loss wt. % (Hebron, F-K)	Hours at temp.	Total loss wt. % (Elba, F-K)	Hours at temp.	Total loss wt. % (Elba, Strunz)	Hours at temp.
50	—	—	—	—	0.04	8
100	—	—	—	—	0.11	8
120	0.01	50	0.00	72	—	—
150	—	—	—	—	0.18	8
160	0.01	26	—	—	—	—
200	0.01	24	0.01	42	0.30	8
225	—	—	—	—	0.49	8
240	0.01	28	—	—	—	—
250	—	—	—	—	0.70	4
280	0.02	48	0.07	42	—	—
300	—	—	—	—	0.86	4
320	—	—	0.10	48	—	—
350	—	—	—	—	1.55	4
360	0.07	40	0.19	46	—	—
400	0.15	27	0.35	77	—	—
440	0.26	48	0.61	67	—	—
480	0.41	50	0.97	74	—	—
500	—	—	—	—	2.58	4
520	0.66	54	1.35	74	—	—
560	0.96	68	1.63	71	—	—
600	1.25	72	1.87	72	—	—
640	1.40	52	1.98	56	—	—
680	1.44	48	2.01	48	—	—
800	1.46	48	2.03	48	—	—
1520	1.46	24	—	—	—	—

The dehydration curves of Fig. 1 seem at first sight to indicate that the water in pollucite is “bound” or “combined” water. Certainly the course of dehydration is not what would be expected of a typical zeolite. Nevertheless, the optical and x-ray studies show that the water may be expelled without appreciably changing the surrounding lattice. The refractive index of the mineral is lowered by dehydration—from 1.523 to

1.513 for the Hebron material, and from 1.520 to 1.506 for the Elba material. This lowering of refractive index is most easily explained by assuming that the water in pollucite occupies interstitial space in the crystal lattice. When the mineral is dehydrated, the water is replaced by air, and the refractive index of the mineral is lowered. The change in refractive index is greater for the Elba pollucite, as would be expected from the higher water content.

Rotation powder spectrum photographs of dehydrated samples of pollucite from Hebron and from Elba were found to be identical with those of the original material and with each other, and also with that obtained from a crystal of pollucite from Greenwood, Maine. This was a portion of the material studied by Richmond and Gonyer (5). The same lines are present in each powder photograph, and no change in the intensities or in the spacings of the lines could be detected.

Changes of the crystal lattice have been observed when some zeolites are dehydrated, but no such effects were observed with pollucite. The length of the cell edge calculated from the measurements of the lines of the powder spectrum photographs is $13.69_6 \pm 0.005 \text{ \AA}$. Previously, others have reported 13.68 (2), 13.71 (6), 13.66 (1), 13.74 (3) and 13.65 (5).

For future identification of pollucite by the *x*-ray powder method, the planar spacings d/n and the relative intensities of the lines are recorded in Table 2. Strunz (6) has recorded intensities of the powder spectrum lines, but his table is incomplete.

According to the *x*-ray work by Strunz (6) and by Naray-Szabo (3), the structure of pollucite is very close to that of analcime. Pollucite is, then, zeolite-like in structure in that the water is not an integral part of the crystal lattice. It differs somewhat from most zeolites in (1) the very low water content, (2) the difficulty of dehydration, and (3) the difficulty of rehydration. This last point is rather interesting. We have kept a sample of pollucite, which had been dehydrated at 500° , in contact with water vapor at room temperature for over six weeks. In agreement with Strunz's result, there was only a very slight gain in weight, which was lost on heating at 110° . No change in refractive index was observed.

It seemed possible, however, that water might be put back into the mineral under pressure. We have carried out experiments in which the dehydrated mineral was heated at $400\text{--}500^\circ$ with water in a bomb. As we had hoped, the pollucite was partly rehydrated by this treatment. It had been thought by us that the mineral might be partly decomposed with the formation of claylike alteration products, similar to those which have been observed at Tin Mountain and at Varuträsk. No such decomposition was observed in our experiments, the pollucite remaining

TABLE 2

Planar spacings and relative intensities of the α -ray diffraction lines of pollucite taken with filtered Cu-K radiation. (These are from our measurements on samples of pollucite from Hebron, Elba and Greenwood, which gave identical values.) The specimen was rotated during the exposure, and a circular camera, calibrated with NaCl, and having a radius of 57.22 mm., was used. Estimated intensities of the diffraction lines are based on a scale of ten, ten being the intensity of the strongest line.

Lines	Intensity	d/n	Indices*
1	2	5.64	112
2	3	4.90	022
3	3	4.17	113
4	4	3.67	123
5	10	3.43	004
6	8	2.925	233
7	2	2.690	134, 015
8	6	2.424	044
9	4	2.224	116, 235
10	1+	2.019	136
11	1+	1.977	444
12	2	1.899	046
13	5	1.863	127, 255, 336
14	1	1.830	246
15	7	1.740	156, 237
16	2	1.713	008
17	1+	1.686	118, 147, 455
18	1	1.637	356
19	2+	1.592	057, 138, 347
20	2+	1.551	257
21	2+	1.531	048
22	2	1.477	129, 167, 556
23	1	1.444	039, 158, 457
24	3+	1.413	239, 367
25	4	1.356	1 1 10, 277
26	3	1.306	1 3 10, 259, 567
27	2	1.282	178, 477, 558
28	2	1.260	169, 3 3 10
29	3+	1.219	1 2 11, 1 5 10, 369
30	3+	1.182	2 3 11, 279, 3 5 10, 677
31	1+	1.164	1 4 11, 578
32	1	1.140	2 6 10
33	3	1.117	1 7 10, 2 5 11, 5 5 10
34	1	1.088	1 6 11, 3 7 10
35	1+	1.062	299, 3 6 11, 679
36	2	1.036	1 2 13, 2 7 11, 5 7 10
37	2	1.016	1 9 10, 2 3 13, 5 6 11
38	2	1.004	1 4 13, 1 8 11, 4 7 11

* Indices based on a cubic lattice.

transparent and isotropic. The refractive index of the material rose with increasing hydration. Recently, Norton (4) has reported a single experiment in which pollucite was heated in a bomb with water and carbon dioxide. No change was observed by him, but as the refractive index of the material was not determined, it is possible that some dehydration occurred in his experiment.

The two most interesting of our experiments, summarized in Table 3, were carried out to test the reversibility of the dehydration. The pro-

TABLE 3. Rehydration experiments

	H ₂ O content wt. %	<i>n</i>
<i>Elba pollucite</i>		
<i>At start of bomb experiment</i>		
Sample A (dehydrated at 500°)	0.28	1.510
Sample B (fresh material) •	2.03	1.520
<i>After heating in bomb at 500° for 24 hrs. with 2 g. H₂O (approx. 300 atm. pressure)</i>		
Sample A	1.29	1.516
Sample B	1.44	1.516
<i>Hebron pollucite*</i>		
<i>At start of bomb experiment</i>		
Sample C (dehydrated at 460°)	0.35	1.516
Sample D (fresh material)	1.46	1.523
<i>After heating in bomb at 493° for 36 hrs. with 3.0 g. H₂O (approx. 380 atm. pressure)</i>		
Sample C	1.12	1.521
Sample D	1.06	1.521

* This experiment was kindly carried out for us by Dr. Earl Ingerson of the Geophysical Laboratory.

cedure used was to place two small platinum crucibles, one containing a known weight of partially dehydrated pollucite, the other containing a known weight of fresh pollucite, side by side in a small bomb with water. After the bomb had been heated at the desired temperature, it was cooled, the crucibles were removed, dried at 110°, and reweighed, and the refractive indices of the samples were determined.

It seems probable from the data in Table 3 that the hydration experiments represent equilibrium conditions. Nevertheless, it is somewhat puzzling that such high pressures are required for partial rehydration of the mineral at 500°. Assuming that the experiments in Table 3 do represent equilibrium conditions, the two samples of pollucite studied require a pressure of at least 400 atmospheres to rehydrate them to their original

condition at 500°. One might reasonably expect that if these samples were placed at 500° under atmospheric pressure, the dehydration would be extremely rapid, even explosive. Actually, dehydration proceeds rather slowly and no decrepitation was noticed.

The rehydrated material does seem to lose its water somewhat more rapidly than the original material. Sample A of Table 3 showed the following loss in weight when heated (24 hours at each temperature): 120° 0.00, 220° 0.06, 300° 0.23, 420° 0.89, 530° 1.29, 620° 1.36 per cent.

The differences in behavior between pollucite and analcime are probably largely due to the fact that the very large cesium ion occupies much of the "free" space in the pollucite structure. This steric effect would account for the low water content and for the difficulty of dehydration. The increase in water content of pollucite as sodium replaces cesium is in accord with this idea. One would expect that the samples with high sodium content would be dehydrated more readily. Unfortunately, none of the samples available to us had over 2 per cent water. Dehydration experiments on material with 3.5–4 per cent water would be of great interest.

ACKNOWLEDGMENT

We are indebted to the late Professor W. E. Ford of Yale University for furnishing the Hebron pollucite from the Brush Collection; to Dr. W. F. Foshag of the U. S. National Museum for supplying the Elba pollucite; and to our colleague Dr. Earl Ingerson for carrying out one of the rehydration experiments.

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ON THE ORIGIN OF SOME PEGMATITES IN THE TOWN OF NEWRY, MAINE

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ABSTRACT

The Newry pegmatites described in this paper include a group of four which are located $6\frac{1}{2}$ miles south of Andover, Maine, in the north-east part of the Town of Newry. The relative locations of the particular pegmatites are shown on a sketch map. The principal pegmatite of the group shows a pronounced banding along the hanging wall; the contact band consists of a chilled deposit of fine-grained albite, near oligoclase. The following band consists chiefly of quartz with some intergrown albite and muscovite. The succeeding bands consist of muscovite and albite (cleavelandite). The repetition of albite constitutes a rhythm and belongs to a type of banding believed to be produced by a process of *rhythmic fractional crystallization*, previously proposed by the author to explain certain banded occurrences in fissure veins.

The origin of the pegmatites is credited to a progressive fractionation of single injections of pegmatite magmas separated in time and space. The injections followed the schistosity of the enclosing rock, except in one case where a soda-lithia pegmatite magma was injected along the foot-wall schist contact of an earlier potash type of pegmatite.

There is no evidence to indicate that any of the magmas contained more than 5 to 10 per cent of water or other volatile constituents. Evidences are also lacking to show that the present soda-lithia pegmatites replaced former potash-rich pegmatites; neither are there evidences of any ingress of later hydrothermal solutions. A later soda-lithia pegmatite in contact with an earlier potash pegmatite did not replace any visible part of the latter.

The minor amount of alteration present can readily be accounted for by the reaction produced by the small amount of residual water and other volatiles which no doubt are liberated as highly heated vapors or liquids, which as a consequence are in a very active state and readily affect an alteration of such minerals as spodumene and triphylite.

The pegmatites of the Newry, Maine, locality are located about $6\frac{1}{2}$ miles south of Andover, Maine, and were described in detail by Fraser (1) in 1930, who concluded (1, p. 363) that:

The pegmatite at Newry fails to show any noteworthy banding or zoning of its constituents. A detailed examination of the mineral sequence, however, shows that there is a rather close parallelism, mineralogically, with the pegmatites at Buckfield and elsewhere, as described by Landes. From the structural and textural data available there appears to be sufficient evidence to show that there were at least three definite periods during the formation of the pegmatite. The first and second periods were transitional but the second and third periods were definitely separated by a period of extensive replacement activity which however, was very local in its action. The evidence indicates that the entire process was essentially continuous.

During August, 1936, the writer spent a few days visiting the pegmatites along the western border of Maine and while at the Newry locality he was impressed by the conspicuous banding, Fig. 3, along the hanging wall of pegmatite No. 2 of Fig. 1, the principal pegmatite of the

group. The most striking band consists of very fine-grained albite (near oligoclase) which occurs next to the hanging wall or is separated from the hanging wall at places by an irregular, interrupted band of black tourmaline.

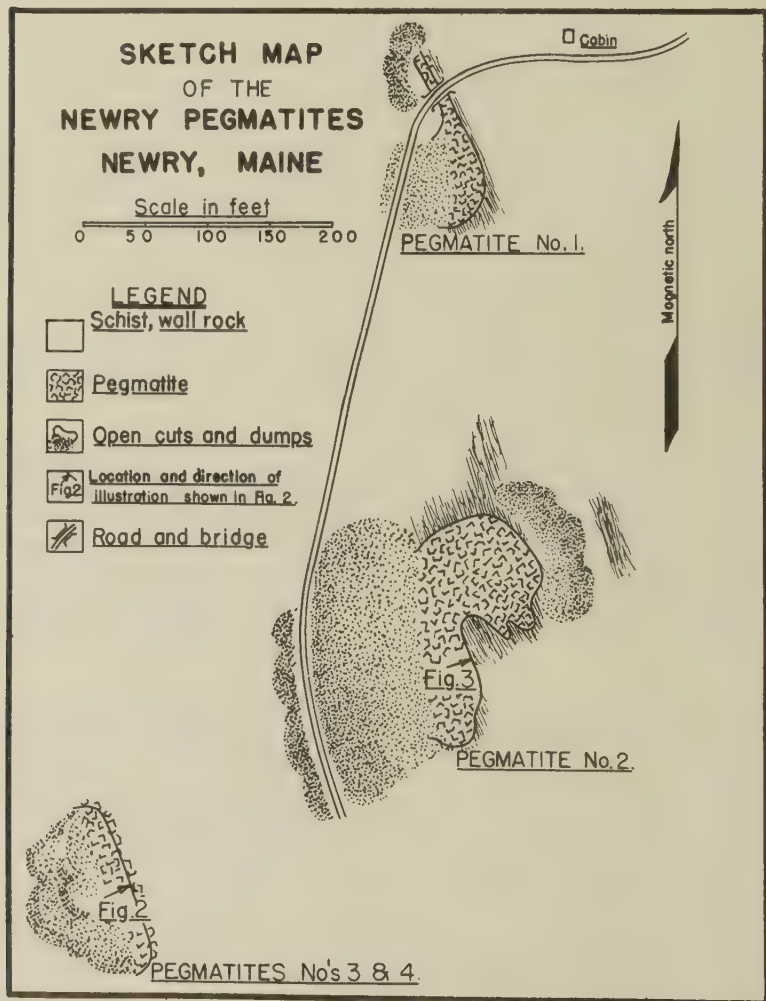


FIG. 1. Sketch map of the Newry pegmatites showing their relative locations.

On account of the unusual occurrence of albite, as a chilled band adjacent to the hanging wall, together with other specific relationships of the minerals which did not appear to harmonize with the paragenesis

given by Fraser (1, p. 352), the pegmatites were revisited on two later occasions.

The Newry pegmatites included in this discussion consist of the two described by Fraser and two additional ones which occur together in a small quarry located several hundred feet to the south and west of pegmatite No. 2. The latter are designated as pegmatites Nos. 3 and 4 on Fig. 1. It could not be definitely determined from Fraser's paper if he described material from pegmatites 3 and 4.

The strike and dip of the pegmatites of the group conform in general to the strike and dip of the schistosity of the enclosing rock. The schist consists essentially of muscovite, biotite and light green actinolite (Fig. 7). In addition it contains a very small amount of quartz which occurs in minute grains very sparsely scattered throughout the rock.

As pegmatite No. 1, Fig. 1, did not appear to contain any specially important evidence on the origin of soda-lithia pegmatites and on the nature of pegmatite magmas, it will not be further considered at this time. It does, however, contain a suite of soda-lithia minerals and it is probably more closely related mineralogically to pegmatite No. 4 than to either No. 2 or No. 3.

Pegmatite No. 2 is most replete with evidences concerning its origin and hence the writer will outline, in more-or-less detail, a theory of origin based on the evidences most clearly revealed by this pegmatite. The banded hanging-wall portion shown in Fig. 3 is well developed, while the contact band of chilled albite is especially noteworthy. Throughout the pegmatite the radial structures developed especially by the cleavelandite, as shown in Figs. 4 and 6, provide abundant examples of the progress of the crystallizing soda-lithia-cesia-rich pegmatite magma.

DESCRIPTION OF THE BANDING ALONG THE HANGING WALL OF PEGMATITE NUMBER 2

The first deposition consists of an irregular, discontinuous crust or band of black tourmaline which is of a deep indigo color in thin section, and which also occurs as "suns," irregular clusters, and as individual crystals in the schist. The latter feature is a common one associated with other pegmatites, whether potash- or soda-rich types, hence it does not appear to offer any critical criteria pertaining to the origin of soda-lithia pegmatites. The first material to be desposited in sufficient quantity to develop a continuous band along the hanging wall, where exposed, consists of a very fine-grained albite, near oligoclase, Fig. 8, which forms a band up to three inches thick. The portion of this band nearer the wall has a more or less equidimensional fine-grained texture. The grains may show many fine or only a few twin lamellae or they may be without ap-



FIG. 2. A portion of the face of pegmatites Nos. 3 and 4. Note the fine-grained contact zone at the base of pegmatite No. 3 and the narrow radial band of muscovite near the top of pegmatite No. 4.

FIG. 3. A portion of the hanging wall section of pegmatite No. 2. The narrow chilled band of albite is clearly shown near the top in contact with the schist. The following band consists largely of quartz which is followed by a band of muscovite. Beneath the muscovite is cleavelandite. The scale across the muscovite band is extended 12 inches.

FIG. 4. A section of a part of two parallel spodumene crystals, containing a band of cleavelandite which grew normal to the adjoining surfaces of the crystals. Like bands of cleavelandite were also attached to the surfaces A and B, but these broke away in handling the specimen.

parent polysynthetic twinning. The shape of the grains near the succeeding band becomes more elongated or "platy," and they begin to show an orientation normal to the band. The second band, up to twelve inches in thickness, consists largely of quartz with intergrown albite and muscovite.

It is not improbable that the second band, with some difference in mineral composition, would have been the first band to crystallize if the wall rock had contained an abundance of quartz so that the quartz could have rendered a "seeding out" effect. Hence, the character of the wall-rock may, under certain physico-chemical conditions, make a decided mineralogical and textural difference in the contact crust.

The third band to form is up to sixteen inches thick and consists largely of long blades of muscovite normal to the wall. The size of the muscovite blades increases with the development of the band, a phenomenon common and probably always present in crystallized mineral crusts (2). Some quartz and cleavelandite crystallized along with the muscovite.

During the crystallization of the preceding albite and quartz bands, very little potassium and water could have been removed from the pegmatite magma by crystallization of the earlier non-hydrous minerals, and the dense, compact, fine-grained albite band must have provided an effective impermeable barrier between the hanging-wall rock and the crystallizing magma. Hence, the water contained in the outer part of the magma was accumulating at the interface between the advancing crusts and the remaining magma. Consequently the composition of the fractionating pegmatite magma became richer in water as well as potash which resulted eventually in a composition and a degree of supersaturation at the interface which caused the precipitation of muscovite. In the absence of much water, other physico-chemical conditions being equal, one may expect that microcline would occur in place of the muscovite band, or it may be deferred to a position nearer the center of the pegmatite mass, depending of course on the amount of K_2O present. The result of a fractionation of this kind may account for the occurrence of microcline nearer the foot wall, as the water of the pegmatite may have been more concentrated toward the hanging wall side. It is probably very rarely that a pegmatite magma has previously undergone sufficient fractionation to eliminate enough potash to prevent the formation of at least some potash feldspar.

The muscovite band was followed by another band of albite; this time the albite developed the characteristic bladed variety known as cleavelandite. At the contact with the preceding band of muscovite the blades of cleavelandite are predominantly smaller, and normal to the earlier bands. The cleavelandite blades increase in size to a maximum as the

distance from the contact increases. As crystallization progressed the temperature conditions became more uniform throughout the pegmatite as the loss of heat was not as rapid and numerous centers of crystallization were initiated by other minerals such as beryl, spodumene, triphylite, tourmaline, etc. From these minerals as centers of attachment the succeeding cleavelandite formed radiating structures as shown in Figs. 4 and 6. The further crystallization of the albite, which constitutes more than half the total quantity of the minerals, followed a random orientation and continued in this manner until crystallization was complete, except for the crusts of albite in the few small cavities where the albite formed a final crust pointing to the center of the cavity.

The banding in contact with the schist is an asymmetrical banding, for that along the foot wall is not a repetition of the banding along the hanging wall. The band along the foot-wall is up to a foot or more thick and consists essentially of medium fine-grained albite. This is followed without sharp demarcation by cleavelandite, of gradually increasing coarseness, which continues into the body of the pegmatite.

Fraser (1, p. 351) states that:

Mr. Butterfield, who visited the quarry in 1927, reported that a layer of microcline about eight inches thick occurred along the foot-wall.

No such band was observed by the writer; however, along with the cleavelandite some distance from the foot-wall a few irregular masses of microcline were observed. It is probable that microcline was more abundant and that considerable commercial microcline may have been removed during mining, however, the principal minerals sought were pollucite, amblygonite, and spodumene.

The visible contacts of the microcline masses with the cleavelandite revealed an intergrowth of the two minerals, as shown in Fig. 5.

It is quite certain that the microcline is considerably later than the first deposition of albite along the foot-wall schist, for as one follows the progress of crystallization away from the fine-grained contact band, or zone of albite, the succeeding albite becomes bladed and has developed blades up to 6 inches or more in length before the microcline began to crystallize. The crystallizing microcline developed around and from the projecting blades of cleavelandite, while at a later time the cleavelandite which continued to crystallize grew away from the microcline when the potash was depleted or reduced below the composition for microcline to continue to form.

The presence of sub- to euhedral crystals of albite within the microcline, Fig. 9, suggests that these were engulfed by the crystallizing microcline and they now represent a poikilitic texture of albite in microcline.

This relationship would be expected from a pegmatite magma in which the vastly more abundant soda-feldspar was simultaneously crystallizing with the microcline.

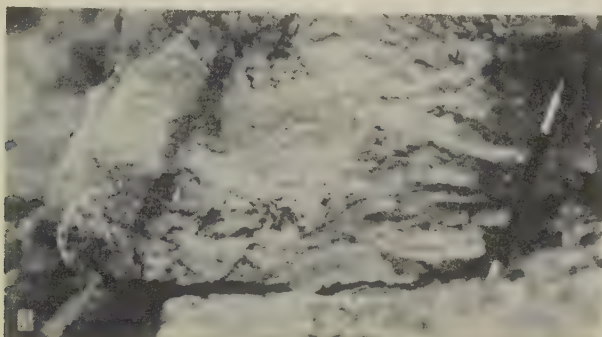


FIG. 5. A specimen of microcline showing, at the right, an intergrowth with cleavelandite.

FIG. 6. A radial growth of cleavelandite about an earlier mass of triphylite, now stained black by manganese oxides. Similar radial growths of cleavelandite are frequent about spodumene, beryl, tourmaline, and other minerals scattered at random within the body of the pegmatite.

The illustration of this intergrowth, Fig. 5, shows oriented blades of cleavelandite on the right intimately associated and intergrown with microcline on the left. This or similar relationships between bladed and massive minerals has been interpreted as being due to the replacement of the massive mineral by the bladed species (3).

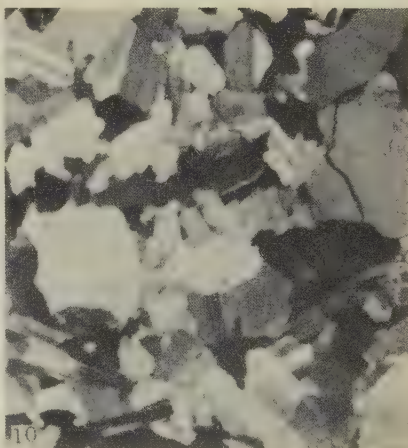
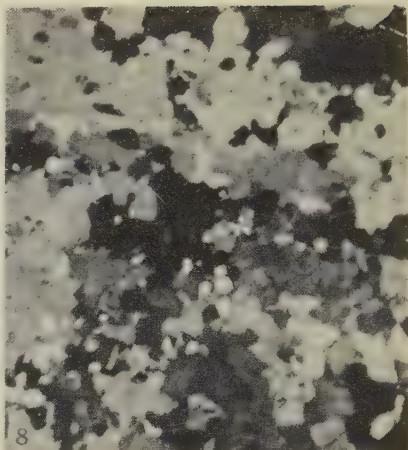
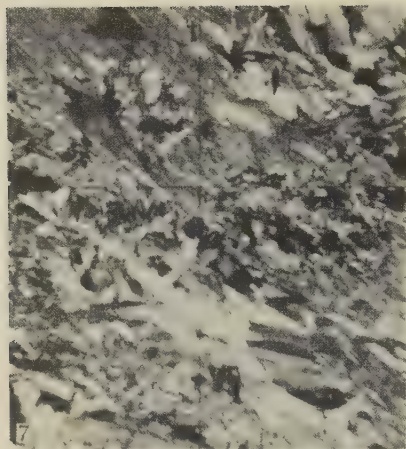


FIG. 7. Section of schist wall rock showing the oriented grains of muscovite, biotite, and actinolite. Quartz is absent in this portion of the section. $\times 16$

FIG. 8. A section of the chilled albite band of Fig. 3 showing the fine-grained texture near the hanging-wall contact. $\times 16$

FIG. 9. Section of microcline showing irregular grains of albite (white) included within the microcline. $\times 16$

FIG. 10. A section of the granular contact rock along the foot-wall of pegmatite No. 3. $\times 16$

RHYTHMIC FRACTIONAL CRYSTALLIZATION

The deposition of albite a second time in the hanging-wall series of bands constitutes a rhythm. The development of the banded structure with its increasingly coarser texture from the wall as described above and shown in Fig. 3 can be explained most logically by the process of

rhythmic fractional crystallization, first proposed by the writer (2, p. 396) to account for the recurring bands of the same mineral species in filled-fissure veins. This occurrence of rhythmic fractionation appears to be the first definite example of the kind observed in a pegmatite and well supported by the mineralogical and textural relationship of the bands and their relationship to the wall rock. Fraser's description of "The Newry Pegmatite" makes frequent references to the recurring deposition of minerals. Among some of his statements are (1, p. 353):

. . . many minor reversals of sequence are to be observed among the Newry minerals; quartz, for example, recurs many times in the general sequence. The recurrence of a mineral later in the general sequence, is further evidence of fluctuations in conditions of the depositing solutions and could hardly be considered as due, in all cases, to the ingress of fresh solutions.

(1, p. 354) Certain minerals, for example, lepidolite or quartz, recurred at numerous periods during the deposition. The frequent recurrence of these minerals probably indicates a local supersaturation of the solution . . .

(1, p. 359) In several specimens a layer of lepidolite was formed around a tourmaline crystal and in one specimen there were *several alternating layers of lepidolite and quartz* around a tourmaline crystal. (Italics are Shaub's.)

This last observation of Fraser's is apparently another excellent example of rhythmic fractionation and undoubtedly occurred rather deep within the body of the pegmatite.

SOME PREVIOUS VIEWS ON ORIENTED MINERAL GROWTHS IN PEGMATITES AND THEIR INTERPRETATIONS

Hess (3, p. 291) states that

much has been made of comb and concentric structures in the solidifying pegmatites. That these exist there is no doubt and they have been explained by crystallizing first along the walls and in successive layers inwards. This feature may be overemphasized, however.

Hess, in the same paper, after thus disposing of the crustified relationships of pegmatite minerals, uses the penetrating relationship between bladed and massive mineral aggregates to establish an earlier crystallization of the latter. The penetrating bladed species are often placed in paragenetic tables as being later than the surrounding or enclosing minerals. In the writer's studies of fissure veins and pegmatites he has found no evidence to support this view, much less the general application of the phenomenon to establish a theory of widespread, extensive and at times almost complete replacement of entire pegmatites. Schaller (4) has also postulated the complete replacement of earlier pegmatites by later solutions thought to be of probable hydrothermal character. Ellsworth (6, p. 329), however, describes an order of crystallization, similar to that proposed by the writer for the No. 2 Newry pegmatite, for some

soda feldspar, microcline, and quartz pegmatites in Ontario. He writes: along the sides next to the wall rock there is a varying width of soda feldspar up to two feet wide along the hanging wall, less along the foot-wall. This is a common, typical arrangement in Ontario pegmatites and would probably be interpreted by most observers as indicating that the soda feldspar crystallized first, the microcline next and finally the quartz.

This brief description suggests a process of crystallization from a pegmatite magma having a composition, except for the fugitive constituents, like the now existing pegmatites, instead of a process in which the soda feldspar replaced an earlier potash feldspar or a graphic-granite which existed between the remaining microcline and wall rock.

PARAGENESIS OF THE MINERALS OF PEGMATITE NO. 2

The order of crystallization as shown in Fig. 11 is based chiefly on observations of material within the quarry and still in place in the pegmatite, or from material on the dumps, or from specimens obtained from the late W. D. Nevel who operated the quarry during the time it produced the many interesting and attractive mineral specimens.

Fraser (1, p. 352) divided the minerals in groups after the grouping of Landes (5). This arrangement seems to the writer to be an unnatural grouping of the minerals of the Newry pegmatites, for he has not found any evidence for making the group distinctions with the exception of the late alteration products which have long been recognized as being secondary. There are variations in the appearance of the same mineral, but such variations are continuous and are to be expected as a pegmatite magma crystallizes with accompanying ever-changing physico-chemical character of the uncrystallized part of the magma still fluid, and continually becoming richer in volatile substances as it reaches the more or less central part of the pegmatite body.

By far the larger quantity of material present in the pegmatite is cleavelandite, and this mineral, with some variation in composition crystallized from the beginning to the end of the period of crystallization, and is certainly magmatic in so far as the originally injected igneous material is magmatic. This long period of crystallization for soda feldspar is not common, even in soda-lithia-cesia-rich pegmatites. Although the soda feldspar may or may not have been slightly preceded at the start by black tourmaline, the total amount of such tourmaline was comparatively small. Quartz and muscovite appeared early and continued in small quantities throughout the consolidation period, the latter becoming richer in lithia as deposition progressed. Microcline also appeared early, but only after a considerable amount of cleavelandite had crystallized.

The minerals, beryl, spodumene, triphylite, and some tourmaline began to appear scattered throughout the pegmatite after approximately a fourth to a third of the magma had solidified, or possibly somewhat earlier. These minerals thus frequently provided a support for cleave-

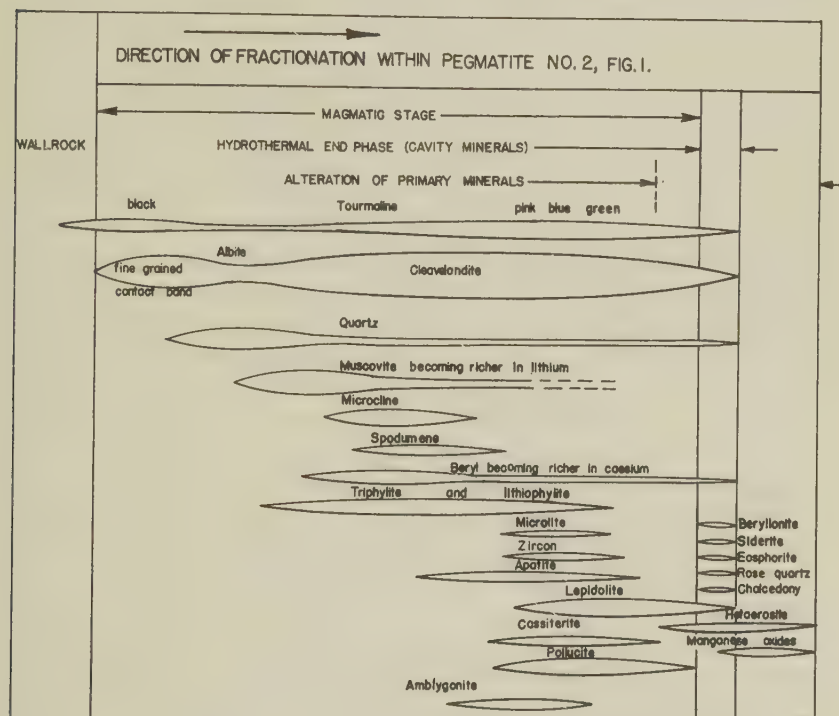


FIG. 11. Diagram showing the paragenetic relationship of the more abundant minerals of pegmatite No. 2.

landite which often produced radial suns about them as centers, as shown in Figs. 4 and 6. Apatite is frequently associated with triphylite as a partial crust and is somewhat later, although close in time of deposition. Lepidolite with characteristic lavender color did not appear until toward the end of crystallization, and is often intimately associated with cleavelandite, while the latter contains, or is associated with, the rarer minerals as microlite, zircon, cassiterite, pollucite, amblygonite, etc., often in a contemporaneous intergrowth. This last association apparently represents the very near-end phase of crystallization when the small amount of water originally present in the magma became more abundant on account of fractionation of the less soluble constituents. The residual water of the pegmatite magma together with mineral matter then in

solution in the water, caused the development of the small crystal-lined cavities or vugs which occur sparingly scattered through a large part of the pegmatite.

The rôle of the small amount of residual water in pegmatite magmas is believed by the writer to be a most important one and to often produce profound effects after some particular phase is reached.

One of the minerals prone to be affected by the residual highly heated water either as a liquid or vapor, or as a phase above a critical point, is spodumene, which in some pegmatites occurs only as completely altered pseudomorphs, or as a core of spodumene with a shell of alteration products. In the Newry pegmatite such profound alteration is absent as the spodumene is comparatively fresh. Another lithium-bearing mineral to be easily affected is triphylite which alters to hetaerosite and manganese oxides. The triphylite throughout the pegmatite has been considerably altered even where all the other minerals appear to be perfectly fresh, Fig. 6. This is probably due to the readiness with which the gaseous water can permeate the cleavelandite of bladed texture and the ease of alteration of the triphylite. As the alteration of the triphylite in the deeper parts of the quarry does not seem to have resulted from surface waters, it appears that the alteration may have started even before the development of the cavity minerals and continued to the present, especially where triphylite is locally exposed to surface conditions.

The textural and mineralogical relationships indicate that the crystallization proceeded within a *closed system*, except for a very brief period at the start due to the porosity of the wall rock, and continued progressively by fractionation from a single injection of a soda-lithia-cesia-rich pegmatite magma. There appears to be no evidence of later solutions of any kind being added to the original pegmatite magma.

DESCRIPTION OF PEGMATITES NOS. 3 AND 4

Pegmatites 3 and 4 occur in the same quarry and to many observers would appear as a single pegmatite. The face of the quarry is illustrated in Fig. 2 which shows a fine-grained portion near the center of the elevation and dipping at a low angle to the right—south. The bottom of this fine-grained band marks the bottom of the granitic contact phase of pegmatite No. 3, and also its contact with the wall rock before the injection of pegmatite No. 4 along the footwall contact. The geologic relationship of pegmatite No. 2 to pegmatites 3 and 4 could not be determined on account of the overburden between the two quarries.

Pegmatite No. 3 is a typical potash-rich pegmatite containing chiefly potash feldspar, black tourmaline, muscovite, and quartz with some coarse graphic granite.

The magma of pegmatite No. 3 was sufficiently superheated to produce a reaction with the country rock forming the granite-like reaction or contact zone, Fig. 10. This contact rock is friable, and difficult to make into thin sections. Three sections of the material analyzed by the Rosiwal method contained an average of 43.9% quartz, 5.8% microcline, 40.2% oligoclase, 8.9% muscovite, 1.2% sphene, 0.1% apatite, and a little black tourmaline.

Following upward from the fine-grained contact rock the pegmatite proper consists of a medium grain, for a pegmatite, which becomes somewhat coarser, but not markedly so. There is some muscovite oriented normal to the contact zone but it is not conspicuous, probably because there is not a large percentage of the mineral present. Also probably because the magma was sufficiently superheated to produce a large mass of highly heated contact rock, as well as to heat the adjoining wall rock to the extent that the heat from the pegmatite proper was not removed sufficiently fast during the period of crystallization to produce marked crustification. Further, water in solution is necessary to produce muscovite.

Pegmatite No. 4 is a more or less typical combination of some potash with abundant soda and lithia minerals. The pegmatite magma was not sufficiently rich in soda to initiate the precipitation of albite, as in pegmatite No. 2. The potash feldspar was first to crystallize in a rather fine-grained deposit in the hanging wall portion attached to the foot wall contact rock of the overlying pegmatite, which had already cooled down sufficiently to somewhat "chill" the magma along the contact of pegmatite No. 4 as it was injected along the fine-grained contact phase of pegmatite No. 3 and the wall rock. A rather pronounced band of muscovite followed the fine-grained border of the lower pegmatite, and this in turn was followed chiefly by coarser grained potash feldspar and quartz. At the foot of the face, the dump material partly covers the central part of the later pegmatite No. 4 which at this point consists of cleavelandite, having rather coarse intersecting blades thus forming openings or "cavities," having polyhedral angles, which frequently contain the rare phosphate eosphorite. Several large crystals of spodumene are visible but these have undergone alteration and are valueless as a source of lithia. The coarse porous nature of the cleavelandite would suggest a greater abundance of residual water than that associated with pegmatite No. 2, and as a consequence a greater hydrothermal attack and leaching, at or near the end of the period of deposition, of the earlier less stable minerals like spodumene.

It is interesting to note that the soda-lithia rich magma of pegmatite No. 4 did *not* react with the earlier overlying potash-rich pegmatite pro-

ducing extensive replacement. Instead, each pegmatite shows clearly that it was derived from a single injection of a pegmatite magma which in all probability, except for the fugitive constituents, had a composition essentially like that of the now existing pegmatites; each having crystallized according to the solubilities of their respective constituent minerals in the residual pegmatite magma as the temperature dropped due to loss of heat to the surrounding rocks.

THE PROBABLE NATURE OF THE PEGMATITE MAGMA

The evidence given above indicates that the magma of pegmatite No. 2 arrived in a single injection and that it had a composition essentially the same as that of the body of the solid pegmatite, except for the escape of a part of the volatile constituents not incorporated in the structure of the minerals during crystallization. The amount of such volatile constituents that escaped is probably of the order of 5 per cent, or even less, rather than that the pegmatite-forming solutions were "hydrothermal." The writer finds too many objections to an origin of the soda-lithia rich pegmatites by a "long continued flow" (3) or by "waves" (5) of hydrothermal solutions.

The temperature of the magma of pegmatite No. 2 was probably not much above the temperature of crystallization of the albite in the contact bands and the wall rocks were also probably at a considerably lower temperature, otherwise the contact bands would not be so fine grained for a soda-lithia pegmatite. The presence of an abundance of water or other gases would have exerted considerable influence on developing a coarse-grained contact and if accompanied by a high degree of superheat, a contact reaction zone with the wall rock would undoubtedly be present as in some other pegmatites.

The absence of a large volume of cavities in the body of the pegmatite shows that any large volume of volatile substances could not have been trapped within the pegmatite magma by the quick consolidation by the peripheral part of the magma into a very dense fine-grained envelope along the contacts.

If one were to follow the paragenesis of Fraser who in turn had followed the grouping of Landes, and placed the soda feldspar later than the microcline, one would have to assume that the early potash feldspar, tourmaline, etc., of Landes' Group I, originally in contact with the walls, had been completely replaced by highly heated aqueous solutions, and then later conditions would have to change sharply to one which would deposit a fine-grained band on the foot wall and a still finer-grained chilled band on the hanging wall. In the interval of dissolving out the preceding pegmatite and the crystallization of the present one, a large

cavity filled with hydrothermal solutions would have had to exist in the schist. As the schist does not show effects of solution, the removal of material would have had to stop at the wall rock, which is inconsistent, for the wall rock should be equally as soluble as muscovite, potash feldspar, quartz, tourmaline, etc., of the supposedly original pegmatite. It is also unlikely that the hanging wall of the schist could have supported itself under such conditions and xenoliths of schist would be expected to be present on the floor in the later soda-lithia pegmatite. If the microcline masses represent residuals of replacement they too should have gravitated to the foot wall and occur only in the foot wall part of the present pegmatite, and they should be surrounded in part, at least, by fine-grained albite just as in the case of the foot-wall schist. Such is not the case for the microcline masses are not anywhere to be seen in contact with the foot wall.

In pegmatites 3 and 4 the conditions of the contacts indicate clearly that the overlying pegmatite No. 3 had already cooled to an appreciable extent before the magma of the lower potash-soda-lithia pegmatite No. 4 was injected. The contact between pegmatites 3 and 4 shows that the material of the later magma did not replace the contact phase of the earlier and overlying pegmatite, neither did it replace any visible part of the earlier potash pegmatite, but, instead a medium fine-grained border or band was deposited and this was followed in places by a band of muscovite having the crystal blades oriented normal to the contact, and they also show a normal directional growth, away from the wall, as crystallization progressed. The relationship of these two pegmatites also shows clearly that in this case at least, and probably in the majority of cases, if the facts were known, the successive pegmatite magma fractions upon being injected into the surrounding schists and gneisses, find a channel of least resistance within the schist or along the contact of the schist and an earlier pegmatite, or other structural weakness rather than within the body of an earlier and consolidated pegmatite. The extremely irregular and lenticular shape of pegmatites in general indicates that the channels through which the pegmatite magmas have passed, undoubtedly have been closed below the large lenticular masses by quick consolidation of the small masses and stringers of magma, that remained within the narrow passages or feeders; for these small masses would freeze very quickly, in general, as is shown frequently by the contact phases of many of the larger pegmatites, hence repeated injections of pegmatite magmas would be expected to produce separate pegmatites of slightly different age and composition.

In a brief statement regarding the origin of the phosphate minerals

from the pegmatites at Poland, Newry, and Buckfield, Maine, and also at Branchville, Connecticut, Berman and Gonyer (7) state that:

This remarkable similarity of four localities at which lithia-pegmatites have been studied seems to require a closed system, with no ingress of new material into the pegmatite solutions.

CONCLUSIONS

From the foregoing description of the structural and textural characters of the minerals of pegmatites Nos. 2, 3 and 4, it seems that these pegmatites originated by the crystallization of single injections of pegmatite magmas. It is certain that the quarry farthest to the southwest in the area is divided into two distinct pegmatites, one overlying and in contact with the other. It is also evident that the soda-lithia-rich magma of the lower one, No. 4, did not in any way react with or replace any visible part of the earlier overlying potash-rich pegmatite.

The evidences found in these pegmatites do not in any way lend support to the replacement theories for the origin of pegmatites rich in soda-lithia- and cesia-bearing minerals.

The writer is indebted to the late W. D. Nevel for many courtesies while studying the Newry pegmatites.

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USE OF THE WULFF NET IN MINERAL DETERMINATION WITH THE UNIVERSAL STAGE

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ABSTRACT

Use of the universal stage in determination of rock-forming minerals, and especially of the plagioclases, is increasing. The plotting of data obtained by such measurements usually presents some difficulty. The Wulff stereographic net is a most convenient and valuable aid in such work for its use obviates much tedious construction. In this paper are given, in the simplest possible form, the necessary procedures for plotting the coordinates of optical symmetry planes, morphological reference planes and their poles. Instructions for transforming the projection to conform with standard reference charts, such as those prepared by Reinhard, are also included.

INTRODUCTION

In recent years use of the universal stage for mineral determination has greatly increased. More particularly, plagioclase determination by the Fedorov method has been recognized as a technique possessing more inherent accuracy than other thin-section procedures. The early Fedorov diagrams¹ were composite in character. They show the migration of the poles of optical symmetry axes in the plagioclase series with respect to conspicuous morphological reference planes. These diagrams are relatively difficult to use and interpret. Simplified stereographic projections of these data were later prepared by Reinhard.² The latter afford a somewhat easier means of plagioclase determination.³ However, plotting and interpretation of the data necessary even for general mineral determination with the universal stage is sometimes inordinately difficult at first. It is the purpose of this paper to describe in detail the essential plotting procedures required in this type of work.

In the writer's opinion, a most essential accessory for plotting the coordinates obtained in mineral determination with the universal stage is the Wulff stereographic net.⁴ With its use much construction with various protractors and scales is eliminated. The net is used directly as a plat, and the coordinates obtained by measurement are quickly and accurately located.

¹ Fedorov, E. S., *Universalmethode und Feldspatstudien*. Part II. Feldspathbestimmungen: *Zeits. Krist.*, **27**, 337-398 (1897).

² Reinhard, M., *Universal Drehtischmethoden*. 119 pp. B. Wepf & Cie, Basel, Switzerland (1931).

³ These five charts are included in Reinhard's book but may also be purchased in separate sets from B. Wepf & Cie, Basel, Switzerland, at a cost of approximately \$0.75.

⁴ Wulff, George, *Untersuchungen im Gebiete der optischen Eigenschaften isomorpher Kristalle: Zeits. Krist.*, **36**, 1-28 (1902).

GENERAL CONCEPTS

In stereographic projection⁵ the center of the crystal, whose faces or optical symmetry planes and vectors are projected, is imagined to be coincident with the center of a *reference sphere* (Fig. 1). By convention,

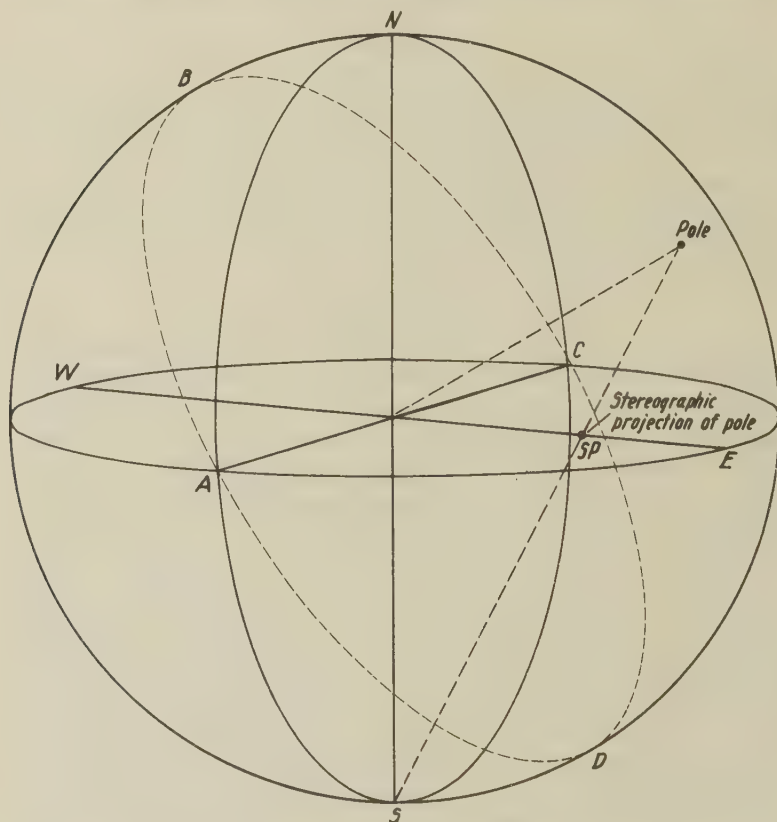


FIG. 1. Reference sphere showing equatorial plane *AWCE*; vertical great circle or meridian *ANCS*; and inclined plane *ABCD*. The pole of *ABCD* and its stereographic projection are also given.

⁵ For additional information on the principles of this projection see:

Penfield, S. L., The stereographic projection and its possibilities from a graphical standpoint: *Am. Jour. Sci.*, ser. 4, **11**, 1-24, 115-144 (1901).

Idem, On the solution of problems in crystallography by means of graphical methods, based upon spherical and plane trigonometry: *Am. Jour. Sci.*, ser. 4, **14**, 249-284 (1902).

Johannsen, A., *Manual of Petrographic Methods*, Chap. II, 2nd. Ed. (1918).

Boecke, H. E., *Die Anwendung der stereographischen Projection bei kristallographischen Untersuchungen*. pp. 58, Gebrüder Borntraeger, Berlin (1911).

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the crystal is oriented so that the crystallographic axis "*c*" coincides with the north-south or vertical axis of the sphere. The eye is assumed to be at the south pole of the sphere. A perpendicular, which passes through the center of the sphere, is erected from the projected face and extended until it intersects the sphere surface. The point of intersection of the face normal with the sphere surface is the *pole* of that face (Fig. 1). A line drawn from any such pole to the south pole passes through the equatorial plane of the reference sphere. The point at which this imaginary line pierces the equatorial plane is the *stereographic projection* of the face or plane in question (Fig. 1). The equatorial plane of the reference sphere is, therefore, the *projection plane* (*Grundkreis*).

In stereographic projection only poles lying in the northern hemisphere, together with those which fall on the equator of the sphere, are customarily projected. Projected poles of all faces parallel to the vertical axis of a crystal lie on the projection circumference. The projected pole of any face which is perpendicular to the vertical axis of the reference sphere, i.e., a horizontal face, lies at the center of the projection.

A stereographic projection has two important properties which make it particularly suitable for crystallographic and universal stage work. First, the projections of both great and small circles are circular arcs and not ellipses. Second, the angular relationship between any two poles on the reference sphere is preserved in the projection. That is, the projection is angle-true (*Winkeltreu*).

DEFINITIONS

A *great circle* (*Grosskreis*) is one with a diameter equal to that of the reference sphere (Fig. 1, *ABCD*). The plane of any great circle passes through the center of the sphere. The projected arcs of great circles pass through diametrically opposite points on the circumference of the projection plane. Any great circle which passes through both the north and south poles of the reference sphere is perpendicular to the equatorial plane and is projected as a straight line. Such perpendicular great circles are called *meridians* or *vertical great circles* (Fig. 1, *ANCS*). Any great circle which makes an angle with the north-south axis appears as a circular arc when projected. The equator of the reference sphere is the only possible horizontal great circle, and its projection coincides with the stereographic projection plane.

A circle with a diameter smaller than that of the reference sphere is a *small circle* (*Kleinkreis*). Horizontal small circles, which lie parallel to the equatorial plane of the reference sphere, are projected as concentric circles about the center of the projection. A vertical small circle is one whose center lies in the equatorial plane and whose plane is parallel to

the vertical axis of the reference sphere. Vertical small circles are projected as circular arcs of varying radii.

DESCRIPTION OF WULFF NET

The base of the standard Wulff net is a circle 20 cm. in diameter (Fig. 2). On this are projected a series of great circles at inclinations of every two degrees with respect to the vertical axis of the reference sphere. At right angles to these, and also every two degrees, is a series of vertical small circles. Since the projected circles are given at two-degree intervals, interpolation with considerable accuracy is possible. Only those parts of the circles lying on the upper hemisphere of the reference sphere are given on the net.

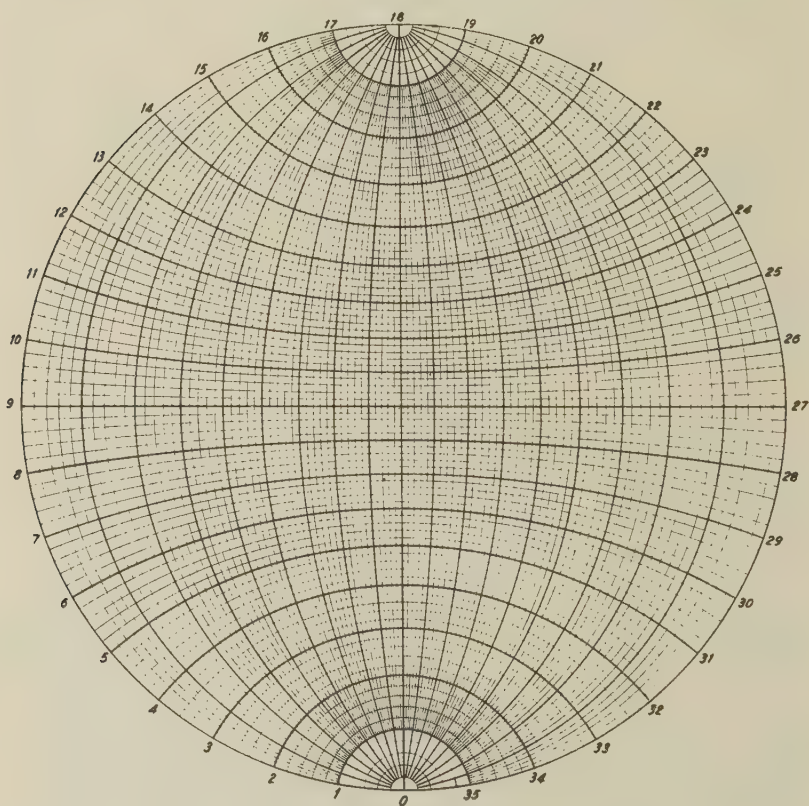


FIG. 2. Wulff net marked at 10° intervals as for use with a standard 4-circle universal stage.

The circumference of the net represents the horizontal great circle corresponding to the equatorial plane of the reference sphere.

The north-south diameter or axis of the net is the projection of a great circle representing a vertical plane containing the north and south poles of the reference sphere. All inclined great circles on the net have the same azimuth. Hence, they pass through the same two diametrically opposite points on the equatorial plane. These two points on the net are the points of intersection of the north-south diameter with the net circumference at which loci the great circles converge.

The east-west diameter of the net is also a straight line. It is a limiting case, being the projection of the largest possible vertical small circle.

When the net is used for plotting the coordinates of a crystal plane, the plane of the thin section is conceived to be coincident with the projection plane. That is, the individual grain examined is visualized as being at the center of the reference sphere, and the planes and axes of optical symmetry are projected therefrom.

With use of the net, projected poles and reference planes of an individual grain can be easily rotated so as to bring any desired plane into coincidence with the projection plane. This is of great assistance in clarifying symmetry problems and in bringing certain optical reference directions into conventional orientation. The ease and rapidity with which planes can be thus interchanged gives the Wulff net a further distinct advantage.

PREPARATION OF NET FOR USE

Wulff nets, commonly printed on heavy paper stock, should be permanently mounted on stiff cardboard, plywood, or bakelite panels. In actual use a sheet of tracing paper, approximately 9×9 inches is a convenient size, is centered over and superposed on the net. All constructions are made by manipulating this sheet; counting off measured inclinations, tracing necessary great circles, and directly plotting poles. If much data are plotted, it is advisable to glue a small piece of tough paper ($\frac{1}{2} \times \frac{1}{2}$ inch) in the center of the sheet. There is then much less chance of enlarging the perforation and introducing errors.

The tracing paper is centered over the net and there fixed, usually with a pin from above. Pins with knurled heads may be procured, but ordinary ones with glass, sealing wax, or cork heads are also quite satisfactory. Alternatively, the center of the net may be pierced from below and the pin or thumb tack fixed with adhesive on the under side.

After superposition of the tracing paper the circumference of the underlying net circle is traced. It is not absolutely necessary to trace the complete circle. Parts of the net circumference only, say small arcs at 120° intervals, may be drawn. This much at least should be done so that the projection may be readily centered, for it will be necessary when making

comparisons with standard reference charts as is often required in plagioclase determination.

Before the net can be used the degrees on its circumference must be plainly marked (Fig. 2). The order in which the degrees are numbered, and the position of the zero point, depend upon the type of stage. With a standard four-circle instrument the 0° mark of the A_1 circle⁶ is directly in front of the operator and nearest him when the stage is in the rest position. Hence, the south end of the north-south diameter of the net is the 0° point. Counting clockwise at the west point is the 90° mark, at the north end is the 180° mark, and at the east point the 270° mark.

If the current Emmons' model 5-circle stage is used, the 0° mark on the A_1 circle is directly opposite and farthest away from the observer when the instrument is in the rest position. The 0° point then lies at the north end of the north-south diameter; the 90° mark at the east point; the 180° mark at the south; and the 270° mark at the west point.

Degrees are usually marked at 10° intervals, and always outside of the net circumference. Two sets of numbers may be laid out so that the net can be used interchangeably with either type of stage without confusion.

An arrow or similar index mark is now made on the circumference of the circle drawn on the tracing paper. This index mark is placed directly over the 0° mark of the net beneath. If proper orientation during both measurement and plotting is to be maintained, this zero point on the tracing paper must be plainly visible.

The coordinates obtained by measurement are often placed in a corner of the tracing paper itself rather than recorded separately. Sketches of crystal outlines, twinning elements, inclusions, and other structures in the measured grain may also conveniently be made directly on the tracing paper. In this way orientation can be verified at a glance and correlation of morphological features, optical planes, and axes previously plotted can be readily made.

CONSTRUCTIONS

The three constructions essential for plotting the customary data secured in mineral determination are as follows:

- (A) Construction of the great circle of a plane.
- (B) Construction of the pole of a great circle.
- (C) Construction of the third plane of optical symmetry.

In addition, it is readily feasible and extremely convenient to

- (D) measure the angle between two stereographically projected planes.
- (E) measure the angle between two stereographically projected poles.

⁶ The axial nomenclature here used is that of Berek. See: Berek, Max, *Mikroskopische Mineralbestimmung mit Hilfe der Universaldrehtischmethoden*, 9-11, Berlin (1924).

To illustrate the essential constructions it will be assumed that the coordinates (Table 1) of two optical symmetry planes and two morphological reference planes in a crystal of plagioclase have been determined.

TABLE 1

Reference Planes	Symbol	Axis		
		A_1	A_2	A_4
Optical symmetry planes				
N_α (constructed)	⊙	209°	56° R(ight)	
N_β	⊥	258°	24° L(ef)t	
N_γ	△	337°	23° R	
Composition plane	◇	136°	13° L	
Cleavage plane	✱	203°	4° L	
Optic axis reading				241°

A. To construct the great circle of a plane

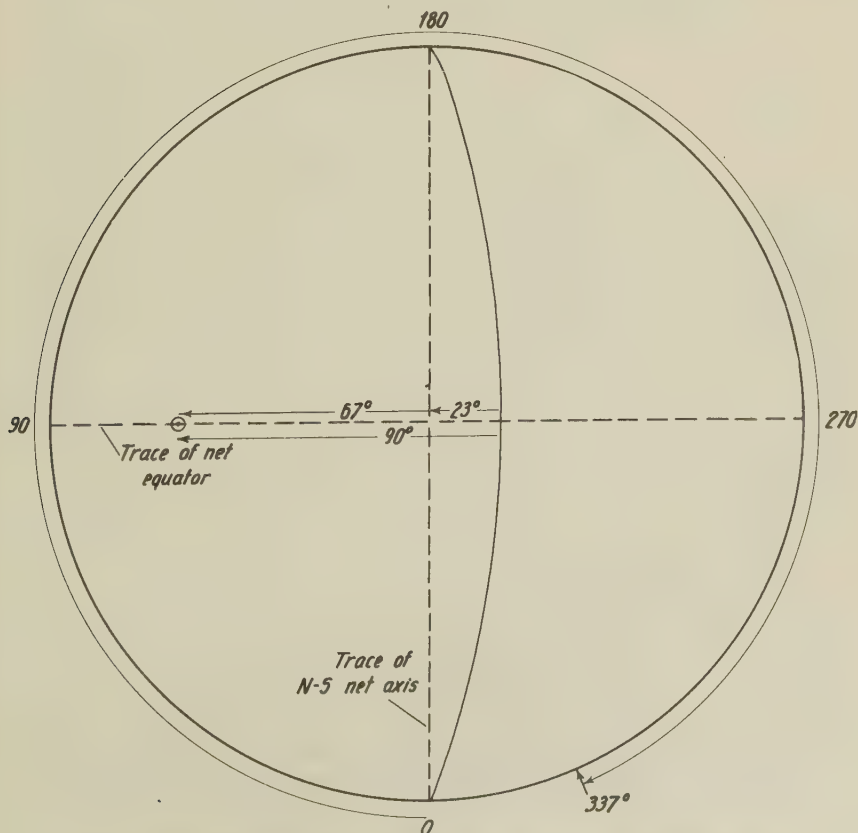


FIG. 3a. Showing the method of plotting a great circle and its corresponding pole.

1. Rotate tracing paper until its index mark rests at that degree mark on the net circumference corresponding to the A_1 circle reading.

2. From the net center, along the equator, count off the number of degrees of rotation about the A_2 axis. (The east-west diameter is usually spoken of as the *equator* of the net.) If the reading is taken off the left arc of A_2 , count left from the net center; if off the right arc, count right from center.

3. On the transparent paper trace the great circle visible on the underlying net, which passes through the point on the equator thus defined.

Example. To plot the plane of N_7 as given in Table 1. Rotate the tracing paper until its index lies at the 337° mark on the net circumference. From the net center, holding the index mark squarely over the 337° point, count to the right 23° along the equator (Fig. 3a). Trace the

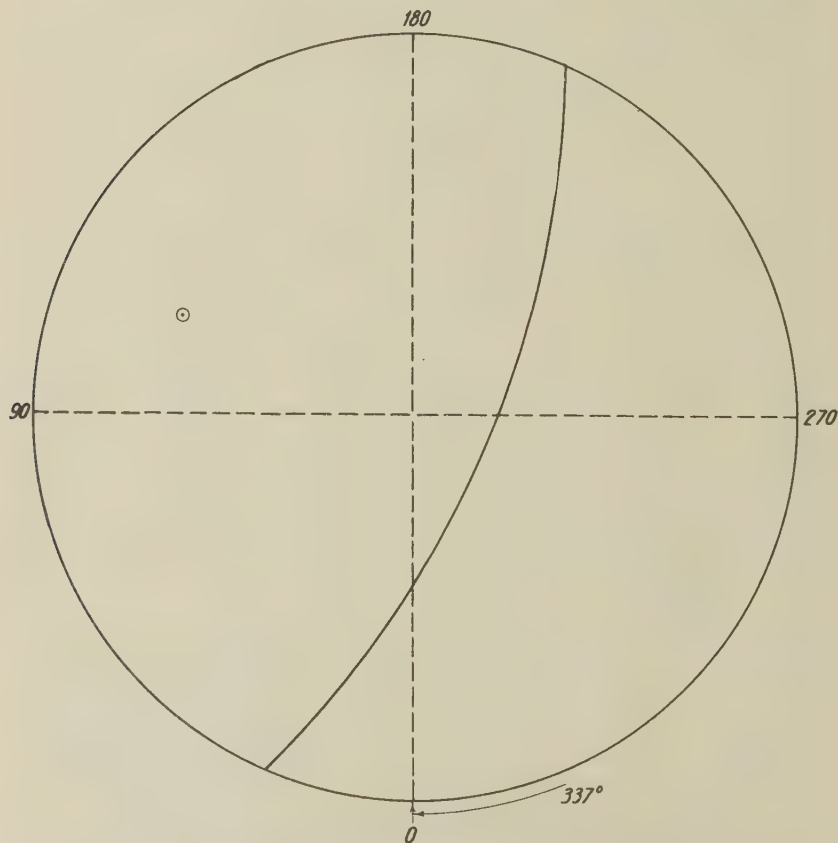


Fig. 3b. The great circle and its pole shown in true orientation after rotation of tracing paper back to 0° on the net circumference = rest position of the stage.

entire arc of the great circle cutting the equator at this point. This arc is the projection of the required great circle. If, now, the tracing paper is rotated back until its index mark lies over the 0° mark on the net, the great circle just drawn will indicate the direction and inclination of the measured plane in space (Fig. 3b).

B. To construct the pole of a great circle

The great circle just constructed represents the stereographic projection of an optical reference plane in the measured crystal. By definition, then, its pole is that point on the reference sphere at which a normal, passing at the same time through the center of the sphere, intersects its surface. The projection of this pole is, therefore, a point 90° from the great circle.

In beginning this construction a plotted great circle must, in all cases, first be brought into coincidence with its corresponding great circle on the net. Hence:

1. Rotate the tracing paper until the index mark lies over that degree mark on the net circumference corresponding to the A_1 reading.
2. Along the equator count 90° to the left or right, depending on the individual case, from the great circle.
3. At the point thus determined prick or mark the tracing paper directly over the net equator. This is the required pole.

In the example the great circle is 23° to the right. Hence, count to the left 67° from the net center and plot the pole (Fig. 3a).

Were the measured plane vertical, its great circle would, when plotted, be a straight line coinciding with the north-south diameter of the net. Its pole would fall on the circumference of the projection. If the measured plane were horizontal, its great circle would coincide with the projection plane and be represented by a circle corresponding to the net circumference. Its pole would be coincident with the center of the projection.

In actual practice a pole is plotted at the same time the great circle is traced. The reverse operation is to draw the great circle of a given pole. To do this the tracing paper is rotated until the pole lies on the equator and a great circle crossing the equator 90° from the pole is then constructed.

C. Construction of the third plane of optical symmetry

In mineral determination with the universal stage the form of the indicatrix and its orientation with respect to crystallographic reference planes are determined. Biaxial minerals possess three mutually perpendicular planes of optical symmetry. To determine the character of the indicatrix the coordinates of two optical symmetry planes are first measured. The great circles and respective poles of these two planes are

plotted. Then, according to the usual methods, the coordinates of the third optical symmetry plane are determined by construction.

Since the three optical symmetry planes are mutually perpendicular, the great circle of any one passes through the pole of both the other planes. Each pole of the two measured planes is 90° distant from P , the intersection point of their respective great circles. The great circle of the third plane is, therefore, that one which joins the poles of the two measured planes. The intersection point P of the great circles representing the two measured planes must also be the pole of the third symmetry plane.

To construct the third plane of optical symmetry:

1. Rotate the tracing paper until the poles of the two measured planes lie on the same great circle of the net (Fig. 4a). If both measurement and previous construction were accurate, point P , at the intersection of the two great circles, will lie on the equator 90° from the great circle defined by the two poles.

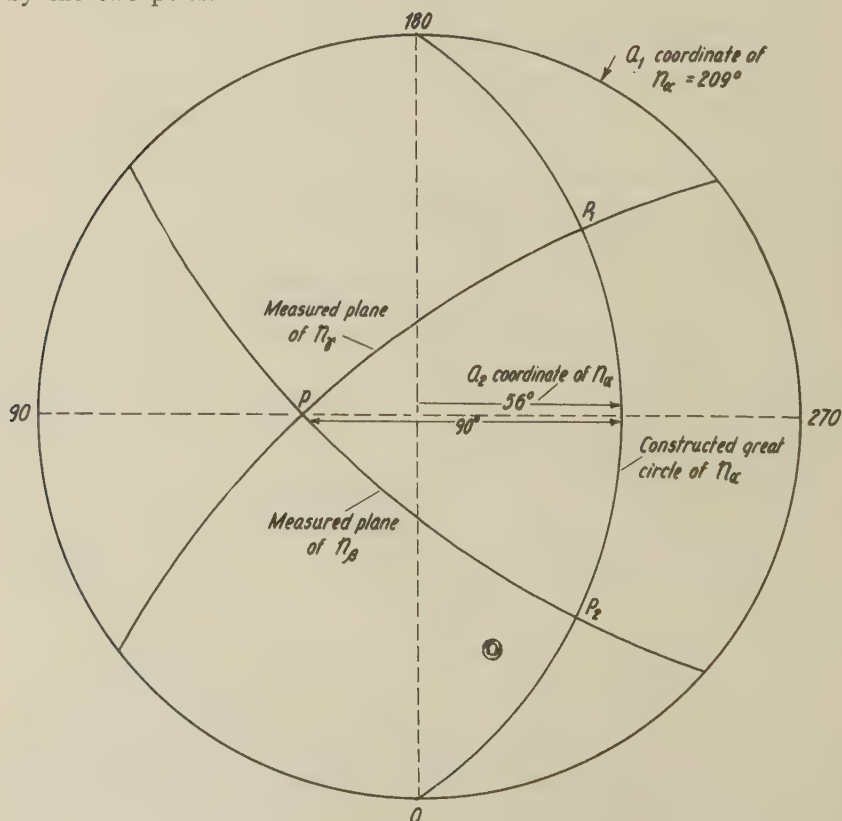


FIG. 4a. Method of constructing the great circle of the third optical symmetry plane and derivation of its coordinates.

2. Trace off the great circle defined by the two poles (Fig. 4a, P_1 and P_2).

3. The coordinates of the third optical symmetry plane can now be obtained.

(a) Hold the tracing paper so that the great circle of the third optical symmetry plane lies exactly over its corresponding great circle on the net. The A_1 coordinate is found on the net circumference at the point where the index mark of the tracing paper then lies.

(b) The A_2 coordinate is found by counting from the center, along the equator, the degrees of inclination of the constructed plane.

In the example the coordinates of the third plane of symmetry were found to be $N_{\alpha}209^{\circ}-56^{\circ} R$ (Fig. 4a).

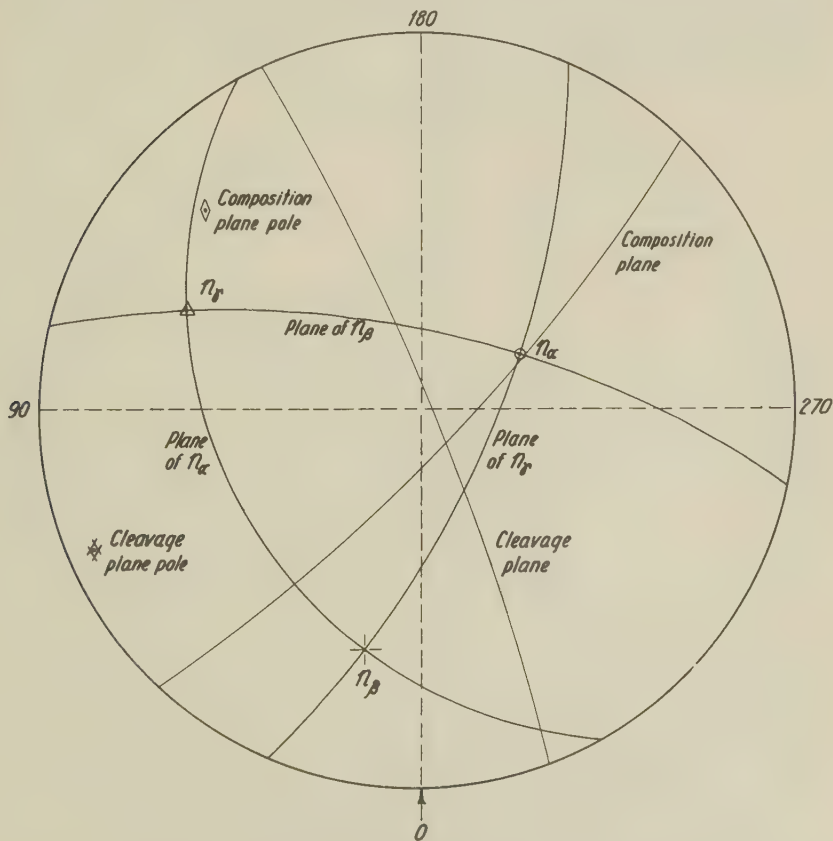


FIG. 4b. Completed stereogram with three optical symmetry planes, composition plane and cleavage plane plotted.

*D. To measure the angle between two
stereographically projected planes*

The true angle between two stereographically projected planes is measured along the arc of the great circle perpendicular to the axis of intersection of the two planes.

1. Vertical planes are projected as straight lines, i.e., as arcs of infinite radii, which intersect at the center of the projection. Since the projection plane is the equatorial plane of the reference sphere, the angle between any two vertical great circles can be measured without further construction.

To find this angle count along the net circumference the degrees of arc between the intersection points R and R' of the two meridians (Fig. 5a). Vertical great circles may be thought of as geographical meridians. Hence, this procedure is tantamount to determining longitude by measurement along the earth's equator.

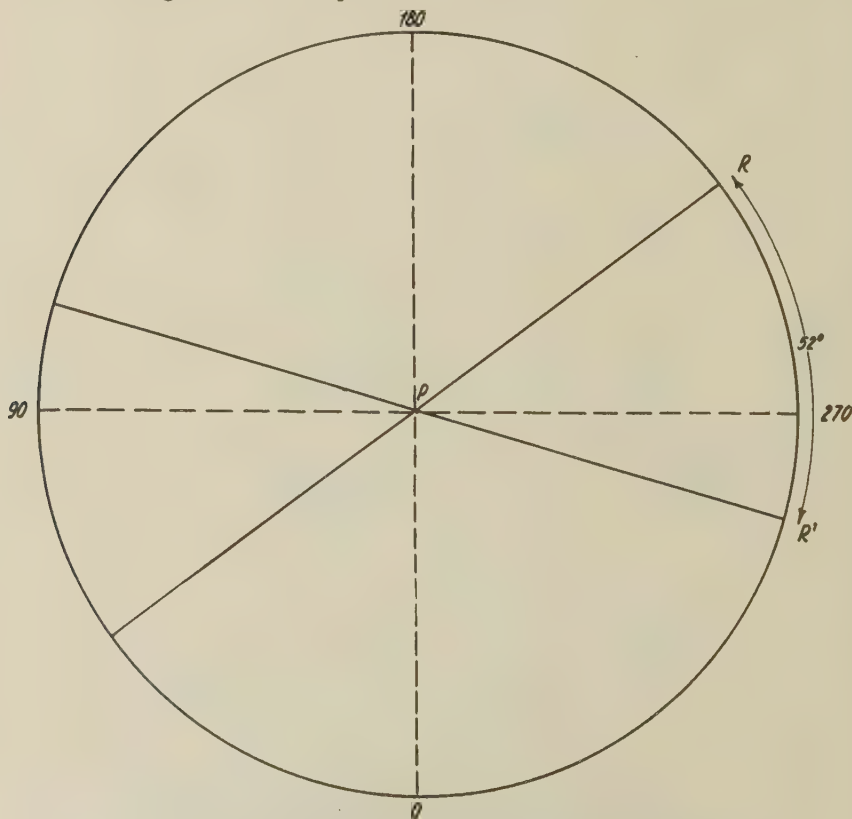


FIG. 5a. Method of determining angle between two vertical planes by measurement along the net circumference.

2. In practice many measured planes will be inclined with respect to the vertical axis of the reference sphere. The great circle of the plane normal to two inclined planes is not coincident with the projection plane. Therefore:

(a) Rotate the tracing paper until P , the intersection point of the two inclined great circles, lies on the equator.

(b) Along the equator count 90° to the left or right of the intersection point P . This will determine the great circle.

(c) Along the arc of this great circle, count the number of degrees between the points of intersection R and R' of the two given great circles. This is the required angle (Fig. 5b).

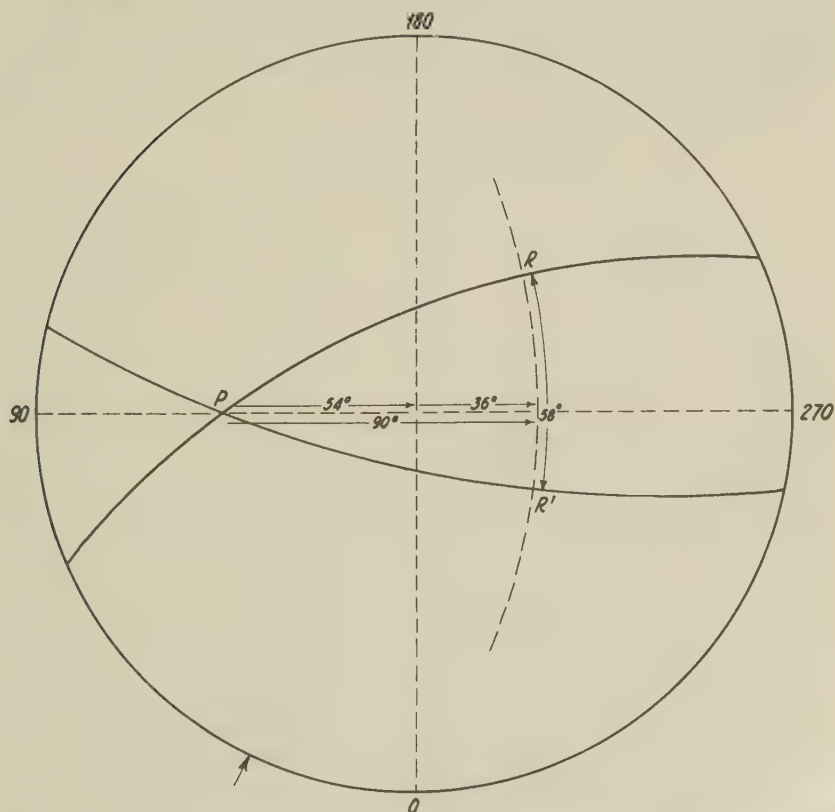


FIG. 5b. Method of determining angle between two inclined planes by measurement along the great circle 90° from their axis of intersection.

The intersection point P of two given great circles may lie on the projection circumference. In this case one plane is horizontal and is, there-

fore, represented by the fundamental circle of the projection. The plane normal to the two given ones is then a vertical great circle. This latter is projected as a straight line passing through the net center. To determine the angular distance between two planes under these circumstances, the tracing paper is rotated until the intersection point P lies on the equator. The required angle is then read off along the north-south axis of the net.

E. To measure the angle between two stereographically projected poles

The angular distance between two poles is measured along the arc of the great circle passing through them.

1. Rotate the tracing paper until both poles lie on the same great circle of the net.

Along the arc of this great circle count the number of degrees between the two poles (Fig. 6).

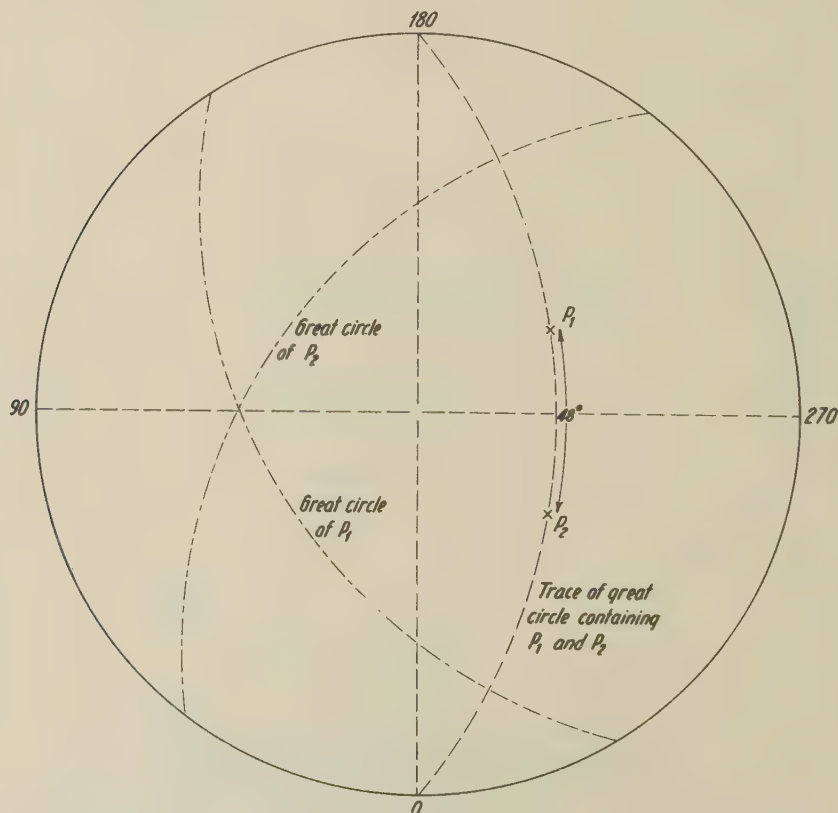


FIG. 6. Illustrating the method of determining the angle between two stereographically projected poles.

If the poles lie on the circumference of the projection, the angle between them is also measured as just indicated. This is so inasmuch as the net circumference represents the horizontal great circle.

In general vertical small circles cannot be used to measure the angle between poles. There is, however, one exception to this. The net equator is the projection of the largest possible vertical small circle and is, therefore, actually a great circle. Hence it may also be used to measure such angular distances.

TRANSFORMATION OF THE PROJECTION PLANE (ROTATION)

It is frequently desirable to bring a particular optical or crystallographic plane in the measured crystal into coincidence with some standard reference plane. Such a procedure is often most helpful in plagioclase determination, especially when the Reinhard method is used. According

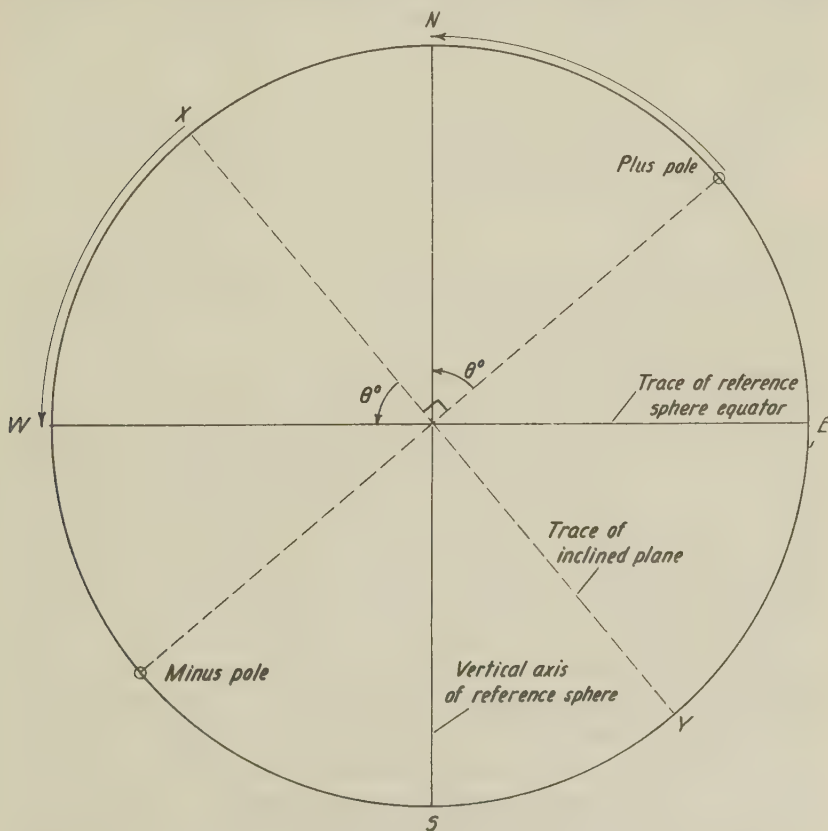


FIG. 7a. Vertical section through reference sphere. Pole rotates through angle θ bringing plane XY into coincidence with projection plane.

to this system the observer's measurements are compared with standard projections of optical vectors and important morphological planes in the plagioclase series. Such projections show the migration of the poles of $N_{\alpha}N_{\beta}N_{\gamma}$, for the plagioclase range, with respect to the common composition planes as (001) and (010).

The great circles on the net, with the exception of the fundamental circle, represent planes which are inclined to the equatorial plane of the reference sphere. All these inclined planes intersect the equatorial plane along one diameter. The line on the net representing this diameter is the north-south meridian, which is also the projection of the only vertical great circle.⁷

To bring an inclined plane, such as $ABCD$ (Fig. 1), into coincidence with the projection plane, the former must be rotated about the diameter AC as an axis. As the plane $ABCD$ is rotated, its pole moves a corresponding distance. It traces a path on the reference sphere surface which, if projected, would appear on the net as part of some vertical small circle (Fig. 7a).

In practice it is easier to rotate a plane by shifting its pole than to manipulate the great circle itself. To transform the projection plane:

1. Rotate the tracing paper until the pole of the plane in question lies on the equator.

With the pole in this position its great circle is superposed over the corresponding great circle on the net beneath. The great circle of the plane in question intersects the equatorial plane of the reference sphere along the proper diameter AC (Fig. 1). If, now, the pole is moved along the equator to the net center, its great circle will then coincide with the projection plane and thereafter be represented by the fundamental circle.

2. Read off on the equator the number of degrees of rotation required to bring the reference pole to the center, and place the pole symbol there.

In rotating the correct angular relationships between the new projection plane and all associated planes must be maintained. To accomplish this the poles of all related planes must be moved in the same direction and the same number of degrees as the pole of the new projection plane. Hence:

3. Shift the poles of all related planes along the arcs of the vertical small circles over which they lie when the pole of the new projection plane lies on the equator as in (1) above (Fig. 7b).

After all required poles are thus rotated and their new positions plotted, their corresponding great circles may be drawn.

⁷ Wulff, George, Über die Vertauschung der Ebene der stereographischen Projection und deren Verwendung: *Zeits. Krist.*, **21**, 248-254 (1892-93).

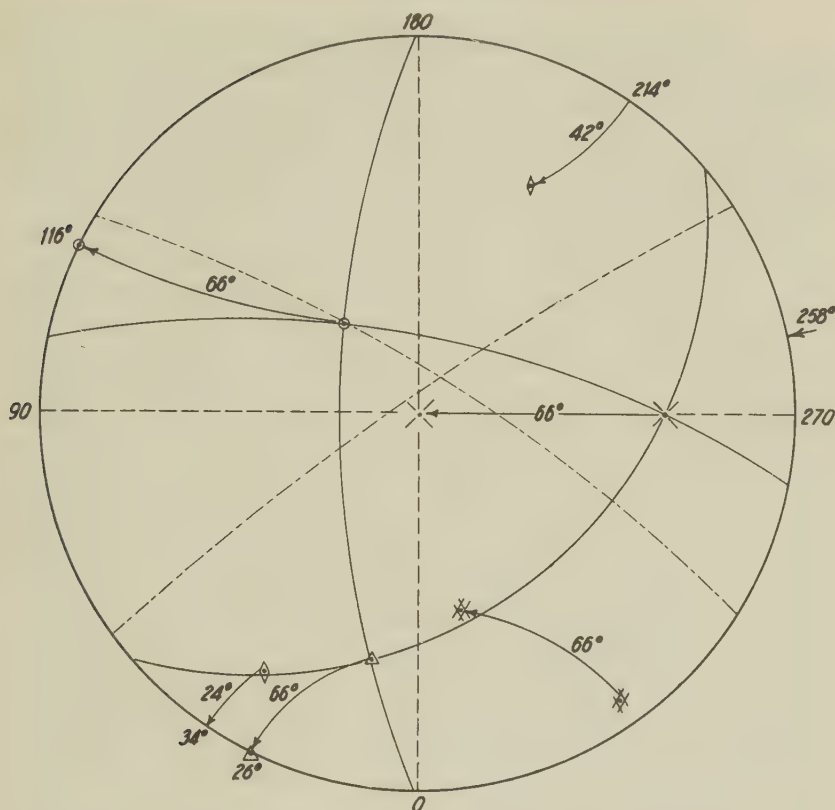


FIG. 7b. Illustrating method of keeping proper orientation by shifting poles of related planes a distance equal to that required to bring reference plane pole to projection center.

1. The tracing paper is rotated so that each newly plotted pole in turn is brought over the net equator.

2. The great circle cutting the equator 90° distant from the pole is then traced.

When several poles representing planes of varying orientation are rotated, some usually lie within the net upon completion of rotation. Often, however, other poles will have arrived at the net circumference before rotation is complete. The proper procedure in this case requires explanation. The normal to any plane, which normal passes through the center of the reference sphere, intersects the lower hemisphere surface at a point diametrically opposite the pole on the upper hemisphere. If during rotation a pole is moved along a vertical small circle beyond the net circumference, it will be carried below the equatorial plane into the lower hemisphere. At the same time the pole of the diametrically opposite extremity of this normal will appear on the upper hemisphere.

As has been noted, in stereographic projection only poles on the upper hemisphere are considered. Let these poles be called *positive* and those on the lower hemisphere *negative*. If a positive pole be rotated to some point on the net circumference, its corresponding negative pole will then lie on the net circumference 180 degrees away. Therefore, having rotated a pole along some vertical small circle until the net circumference is reached:

1a. Count 180° in either direction around the circumference. The point of emergence of the negative pole lies on the point thus determined.

1b. Or, alternatively, a line may be drawn passing through the net center and the point on the circumference where the positive pole lies.

2. After the point of emergence of the negative pole is located, the latter is rotated the remainder of the required angular distance along the arc of that small circle intersecting the circumference at that point (observe composition plane pole, Fig. 7a).

The coordinates previously given in Table 1 will again be used for illustration. For comparison with standard projections it may be necessary to transform a projection so that the optic axial plane of the measured crystal becomes the projection plane. If this is done the pole of the optic normal will lie at the center of the projection. In Fig. 7b the three optical symmetry planes, the cleavage plane, and the twinning plane, together with their respective poles, have all been plotted. The tracing paper has been rotated until the pole of N_β lies on the equator 66° to the right of center. In this position the index arrow lies on the 258° mark on the net circumference. The poles of N_α and N_γ lie on the same great circle 24° to the left of center.

The pole of N_β must now be moved along the equator 66° to the net center. This will bring the optic axial plane into coincidence with the projection plane. The poles of N_α and N_γ must, likewise, be shifted to the left 66° along their respective vertical small circles. They will then lie on the net circumference. The pole of N_α comes to rest at the 116° point; that of N_γ at the 26° mark on the circumference. They are, it should be noted, 90° apart. The composition plane pole is now moved along its appropriate small circle 66° as indicated. After rotation of 24° it reaches the circumference at the 34° point. It is then moved 42° farther along the corresponding small circle, intersecting the net circumference 180° from the 34° point. This diametrically opposite point is at the 214° mark. After the pole of the cleavage plane is rotated the entire projection is transformed.

If it is required to draw the great circles of these poles in their new positions, the tracing paper is turned until each relocated pole in succession lies on the equator. As before each great circle is drawn by tracing the great circle cutting the equator 90° distant from the pole.

The great circles of N_α and N_γ are now meridians making an angle of 90° with one another. The great circle of the cleavage plane lies 34° to the left of center; that of the composition plane 22° left of center (Fig. 8).

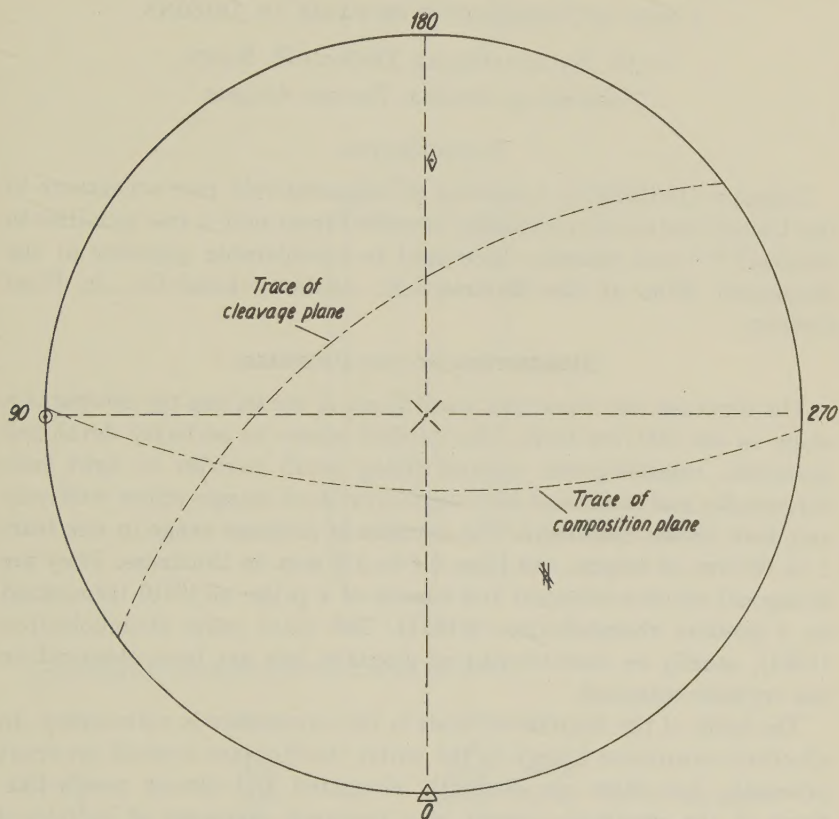


FIG. 8. The transformed stereogram with the optic axial plane coincident with the projection plane.

The stereogram is then completely transformed. The optic axial plane is now the projection plane and all morphological reference planes are in their proper relationship, both with respect to one another and to the planes of optical symmetry.

ACKNOWLEDGMENT

The writer wishes particularly to thank Mr. Martin Capp, of the Civil Engineering Department, Colorado School of Mines, for his generous assistance in preparation of the figures.

NOTES AND NEWS

A NEW OCCURRENCE OF DIOPTASE IN ARIZONA

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INTRODUCTION

Dioptase (H_2CuSiO_4), a mineral of comparatively rare occurrence in the United States and previously described from only a few localities in Arizona,^{1,2,3,4} was recently discovered in considerable quantity at the Mammoth Mine of the Mammoth-St. Anthony Lead Co., in Pinal County.

DESCRIPTION OF THE DIOPTASE

The diopase was found in a small block of ore during the mining of a stope on the 500-foot level. The mineral occurs as perfectly developed prismatic, emerald-green crystals lining small cavities in light blue chrysocolla, and associated with small crystals of orange-yellow wulfenite and dark brown descloizite. The crystals of diopase range in size from 2 to 10 mm. in length, and from 0.1 to 1.5 mm. in thickness. They are hexagonal tri-rhombohedral and consist of a prism $m(10\bar{1}0)$ terminated by a positive rhombohedron $s(11\bar{2}1)$. The third order rhombohedron $(13\bar{4}1)$, usually so characteristic of diopase, has not been observed on the crystals examined.

The habit of the diopase crystals in this occurrence is noteworthy. In all other occurrences known to the writer the diopase crystals are short prismatic, but these are distinctly elongated and almost needle-like. Some of the specimens consist of a confused aggregate of individual prismatic crystals; in others the crystals form small sheaf-like bundles. Some of the diopase projects from the faces of small tabular wulfenite crystals, and tiny crystals of descloizite are perched on the diopase. One particularly fine group of diopase is completely encrusted with black, earthy manganese oxide.

¹ Hills, R. C., Reported occurrence of diopase: *Am. Jour. Sci.*, 3rd series, **23**, 325 (1882).

² Smith, W. B., Dioptase from Pinal County, Arizona: *Proc. Colo. Sci. Soc.*, **2**, 159 (1887).

³ Lindgren, Waldemar, The copper deposits of the Clifton-Morenci district, Arizona: *U. S. Geol. Survey, Prof. Paper* **43**, 111 (1905).

⁴ Lausen, Carl, and Wilson, E. D., Gold and copper deposits near Payson, Arizona: *Univ. Ariz., Ariz. Bur. Mines, Bull.* **120**, 27 (1925).

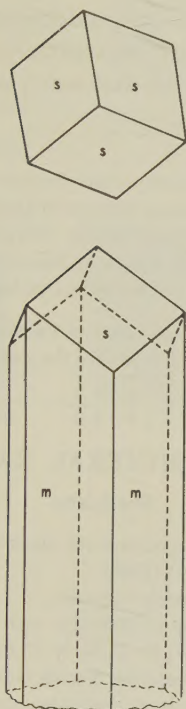


FIG. 1. Diopside.

Goniometric measurements are as follows:

No.	Form	Measured	
		ϕ	ρ
1	10 $\bar{1}$ 0	0°00'	90°00'
2	11 $\bar{2}$ 1	30°00'	51°02' (50°48' calculated)

Optically the diopside is uniaxial positive with $\omega=1.654$ and $\epsilon=1.708$.

Microscopic examination of the chrysocolla revealed the presence of isolated and fan-shaped aggregates of colorless to very pale blue slender crystals about 0.1 mm. in length. These crystals are biaxial negative with high birefringence. Indices of refraction are $\alpha=1.692$, $\beta=1.725 \pm .003$, $\gamma=1.725 \pm .003$. $2V=45^\circ \pm$. These properties correspond most closely to those of tyrolite, as described by Larsen and Berman,⁵ but the crystals appear to be nonpleochroic and microchemical tests failed to reveal the presence of arsenic. In spite of careful search sufficient material could not be isolated for an x-ray diffraction pattern.

⁵ Larsen, E. S., and Berman, Harry, The Microscopic Determination of the Non-opaque minerals: *U. S. Geol. Survey, Bull.* **848**, 194 (1934).

The writers wish to express their indebtedness to Dr. Harry Berman, Department of Mineralogy and Petrography, Harvard University, who kindly criticized the manuscript, and made valuable suggestions.

Dr. Edward Wigglesworth, for twenty years director of the Boston Society of Natural History, has resigned to become Eastern director of the Gemological Institute of America and chairman of its educational advisory board. The gemological laboratory at 69 Newbury Street, Boston, will be under his direction. This is the second gemological laboratory of the institute, the first having been conducted in Los Angeles since 1933.

Mr. Edward W. Nuffield of the University of Vancouver has been appointed Teaching Fellow in Mineralogy at Stanford University for the year 1940-41.

NEW MINERAL NAMES

Norilskite

O. E. ZVIAGINCEV, New mineral species of the platinum group. *Compt. Rend. (Doklady) Acad. Sci. U.R.S.S.* **26**, No. 8, 788-791 (1940).

NAME: From the Norilsk copper-nickel deposits.

CHEMICAL PROPERTIES: An alloy of platinum with iron, nickel and copper. Analysis: Insol. 0.40, S 0.71, Pt. 35.50, Pd 3.57, Fe 25.30, Ni 25.64, Cu 9.28. Sum 100.40.

OCCURRENCE: Found in placers near the Norilsk copper-nickel sulfide ore deposits with grains of other palladium-platinum alloys containing variable amounts of Fe, Ni and Cu.

W. F. FOSHAG